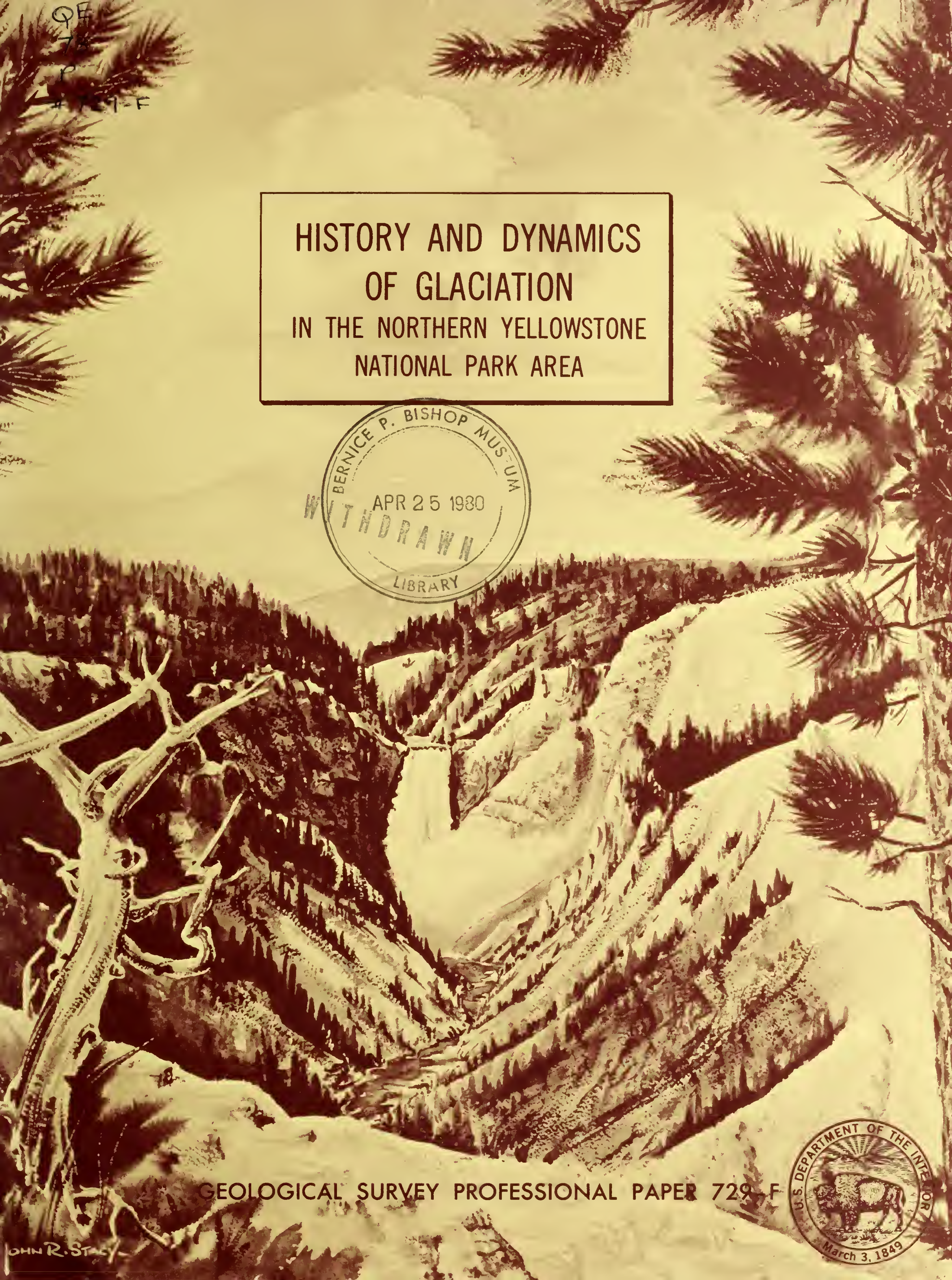
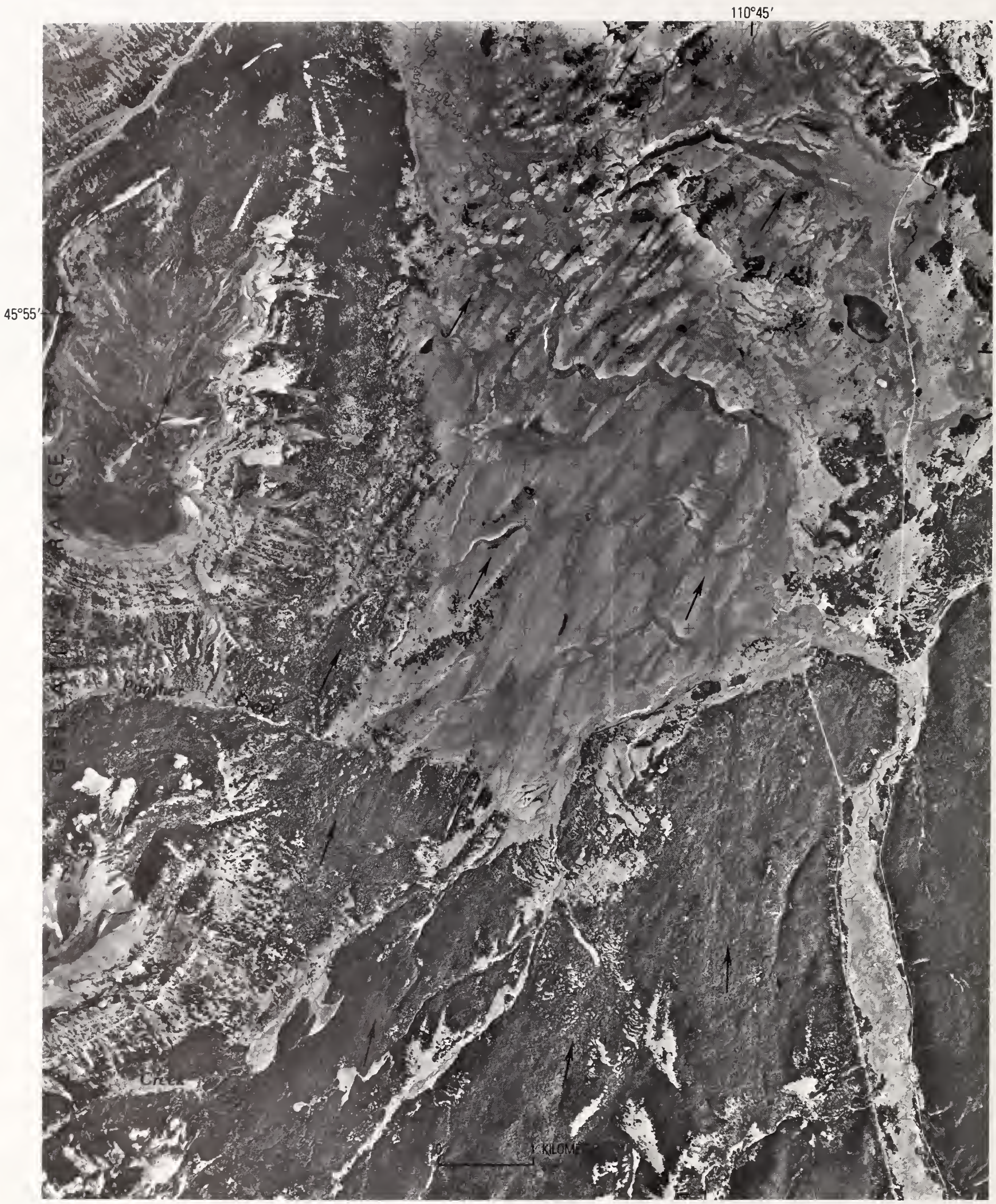
HISTORY AND DYNAMICS OF GLACIATION IN THE NORTHERN YELLOWSTONE NATIONAL PARK AREA



GEOLOGICAL SURVEY PROFESSIONAL PAPER 729-F

HISTORY AND DYNAMICS OF
GLACIATION IN THE NORTHERN
YELLOWSTONE NATIONAL PARK AREA



FRONTISPIECE.—Glaciation of Yellowstone National Park was more extensive and involved more interaction of ice from multiple sources than previously recognized. The drumloid topography (arrows) was formed by ice about 600 m thick flowing oblique to the Gallatin Range front; it shows that glacial flow from the valleys in the Gallatin Range was diverted northward by a thick icecap on the plateau east of the range. Thus the northern Yellowstone glaciation must be considered in terms of coalescing icecaps that covered almost the entire landscape rather than just as valley and piedmont glaciers emanating from the main ranges. Aerial photograph by NASA (National Aeronautics and Space Administration), 1969.

History and Dynamics of Glaciation in the Northern Yellowstone National Park Area

By KENNETH L. PIERCE

GEOLOGY OF YELLOWSTONE NATIONAL PARK

GEOLOGICAL SURVEY PROFESSIONAL PAPER 729-F

A reconstruction and quantitative evaluation of the interacting icemasses that covered northern Yellowstone during the Pinedale Glaciation and the contrast with the preceding Bull Lake Glaciation

U.S. DEPARTMENT OF THE INTERIOR
March 3, 1849

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1979

UNITED STATES DEPARTMENT OF THE INTERIOR

CECIL D. ANDRUS, *Secretary*

GEOLOGICAL SURVEY

H. William Menard, *Director*

Library of Congress Cataloging in Publication Data
Pierce, Kenneth Lee.
History and dynamics of glaciation in the northern Yellowstone National Park area.
(Geology of Yellowstone National Park)
(Geological Survey Professional Paper 729–F)
Bibliography: p. 84
1. Glacial epoch–Yellowstone National Park. 2. Glacial landforms–Yellowstone National Park.
3. Yellowstone National Park. I. Title. II. Series. III. Series: United States Geological Survey
Professional Paper 729–F.
QE697.P64 551.3'1'0978752 78-18782

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C. 20402
Stock Number 024-001-03268-6

Yellowstone National Park, the oldest of the areas set aside as part of the national park system, lies amidst the Rocky Mountains in northwestern Wyoming and adjacent parts of Montana and Idaho. Embracing large, diverse, and complex geologic features, the park is in an area that is critical to the interpretation of many significant regional geologic problems. In order to provide basic data bearing on these problems, the U.S. Geological Survey in 1965 initiated a broad program of comprehensive geologic and geophysical investigations within in the park. This program was carried out with the cooperation of the National Park Service, and was also aided by the National Aeronautics and Space Administration, which supported the gathering of geologic information needed in testing and in interpreting results from various remote sensing devices. This professional paper chapter is one of a series of technical geologic reports resulting from these investigations.

CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

SI UNITS-U.S. CUSTOMARY EQUIVALENTS

[SI, International System of Units, a modernized metric system of measurement]

SI unit		U.S. customary equivalent	
Length			
millimeter (mm)	=	0.039 37	inch (in)
meter (m)	=	3.281	feet (ft)
	=	1.094	yards (yd)
kilometer (km)	=	0.621 4	mile (mi)
	=	0.540 0	mile, nautical (nmi)
Area			
centimeter² (cm²)	=	0.155 0	inch² (in²)
meter² (m²)	=	10.76	feet² (ft²)
	=	1.196	yards² (yd²)
	=	0.000 247 1	acre
hectometer³ (hm²)	=	2.471	acres
	=	0.003 861	section (640 acres or 1 mi²)
kilometer² (km²)	=	0.386 1	mile² (mi²)
Volume			
centimeter³ (cm³)	=	0.061 02	inch³ (in³)
decimeter³ (dm³)	=	61.02	inches³ (in³)
	=	2.113	pints (pt)
	=	1.057	quarts (qt)
	=	0.264 2	gallon (gal)
	=	0.035 31	foot³ (ft³)
meter³ (m³)	=	35.31	feet³ (ft³)
	=	1.308	yards³ (yd³)
	=	264.2	gallons (gal)
	=	6.290	barrels (bbl) (petroleum, 1 bbl=42 gal)
	=	0.000 810 7	acre-foot (acre-ft)
hectometer³ (hm³)	=	810.7	acre-feet (acre-ft)
kilometer³ (km³)	=	0.239 9	mile³ (mi³)
Volume per unit time (includes flow)			
decimeter³ per second (dm³/s)	=	0.035 31	foot³ per second (ft³/s)
	=	2.119	feet³ per minute (ft³/min)
Volume per unit time (includes flow)—Continued			
decimeter³ per second (dm³/s)	=	15.85	gallons per minute (gal/min)
	=	543.4	barrels per day (bbl/d) (petroleum, 1 bbl=42 gal)
meter³ per second (m³/s)	=	35.31	feet³ per second (ft³/s)
	=	15 850	gallons per minute (gal/min)
Mass			
gram (g)	=	0.035 27	ounce avoirdupois (oz avdp)
kilogram (kg)	=	2.205	pounds avoirdupois (lb avdp)
megagram (Mg)	=	1.102	tons, short (2 000 lb)
	=	0.984 2	ton, long (2 240 lb)
Mass per unit volume (includes density)			
kilogram per meter³ (kg/m³)	=	0.062 43	pound per foot³ (lb/ft³)
Pressure			
kilopascal (kPa)	=	0.145 0	pound-force per inch² (lbf/in²)
	=	0.009 869	atmosphere, standard (atm)
	=	0.01	bar
	=	0.296 1	inch of mercury at 60°F (in Hg)
Temperature			
temp kelvin (K)	=	[temp deg Fahrenheit (°F) +459.67]/1.8	
temp deg Celsius (°C)	=	[temp deg Fahrenheit (°F) −32]/1.8	

STATEMENT ON MEASUREMENT UNITS AND STRATIGRAPHIC NOMENCLATURE

Measurements made during field work for this report were in U.S. Customary units. With the exception of altitudes, all values have either been converted or remeasured, and are reported in metric (SI) units, rounded off to express the appropriate degree of accuracy. Where the measurement is of altitude, the altitude in feet is also given in parentheses following the value in SI units, because the base-map contours are in feet. Thus if I refer to glacial striations at an altitude of 2,750 m (9,000 ft) on Specimen Ridge, the altitude in feet is given so the reader can readily locate the feature on the map (pl. 1*A*) using the base-map contours. The contours on the reconstructed ice surface (pl. 1*A*) are in U.S. Customary units because they are derived directly from the topographic base maps. This allows the reader to readily determine the control for the ice-surface contours, as well as their accuracy, by examining them in conjunction with the altitudes shown on the base map.

If this study were initiated in 1979, rather than 1965, I would recommend defining new stratigraphic names for the local surficial geologic units in the northern Yellowstone area. I would recommend this because deposits of the Bull Lake and Pinedale Glaciations cannot be traced by continuous mapping or by lithologic characteristics from their type areas on the flanks of the Wind River Mountains to northern Yellowstone. Also, compared to the Wind River area, glacial deposits in the northern Yellowstone area are constructed from different rock types, have different particle size distributions, and have experienced a generally wetter climate, making interpretation of relative-age parameters uncertain. Nevertheless, the terms Bull Lake and Pinedale were already in use in Yellowstone at the start of this study, and these terms have been used in more than 25 maps and reports published as a result of the U.S. Geological Survey study of the entire Yellowstone area. In spite of my misgivings about the appropriateness of using these stratigraphic terms in northern Yellowstone, I do so in this concluding report because of their use in earlier published reports of this study, and because introduction of new names at this time would require extensive revisions and additions to this report.

HISTORY AND DYNAMICS OF GLACIATION IN THE NORTHERN YELLOWSTONE NATIONAL PARK AREA

By Kenneth L. Pierce

ABSTRACT

Large masses of ice formed in northern Yellowstone National Park during the Bull Lake and Pinedale Glaciations. Bull Lake terminal moraines of the northern Yellowstone icemass occur only on the park's western margin, in the West Yellowstone area. The Bull Lake Glaciation here is about 150,000 years old as dated by combined obsidian hydration and K-Ar methods. Bull Lake ice advanced into the West Yellowstone basin largely through lowlands later filled by the West Yellowstone rhyolite flow. The moraines deposited by this advance are bulky, loess mantled, and have the muted but clearly glacial constructional morphology typical of the Bull Lake elsewhere in the Rocky Mountains. Bull Lake glaciers extended 20 kilometers beyond Pinedale glaciers west of the park; but north of the park, Pinedale glaciers overrode and obliterated Bull Lake terminal moraines. This difference is attributed to emplacement of large rhyolite flows in the western part of the park in the interval between the Bull Lake and the Pinedale maxima. These rhyolite flows have K-Ar ages between 70,000 and 120,000 years old. They did not affect the flow of Bull Lake ice but diverted Pinedale flow by blocking westward flow and shunting it to the north.

Just downvalley from the Pinedale terminus of the outlet glacier in the Yellowstone valley are moraines of both Bull Lake and Pinedale age deposited by *local* glaciers. The location and degree of preservation of these moraines demonstrate that if any outlet glacier existed larger than the Pinedale one, it was of pre-Bull Lake age.

At some time between the Pinedale and Bull Lake glacial maxima, ice from the Beartooth uplift advanced southward beyond the Washburn Range and dammed the Yellowstone River, leaving erratics of Precambrian crystalline rocks.

During Pinedale time, major ice streams from four sources converged near Gardiner, Mont., to form the northern Yellowstone outlet glacier. This outlet glacier then flowed 60 km down the Yellowstone valley to the Eightmile terminal moraines, and although it was joined by several tributary glaciers along the way, the main outlet glacier dominated ice flow.

The northern Yellowstone outlet glacier and its accumulation area together constituted the northern Yellowstone glacier. This glacier covered 3,400 km^2, averaged more than 700 m in thickness, and included a flow line 146 km long. Five interconnected icecap sources contributed to the northern Yellowstone glacier: an icecap on the Gallatin Range, an icecap on the plateau between the Gallatin and Washburn Ranges, an icecap on the plateau south of the Washburn Range, an icecap on the Beartooth uplift, and an icecap in the upper Lamar drainage basin (numbered I–V in fig. 5).

The ice divides along the crest of these icecaps defined the outer boundary of the northern Yellowstone glacier. Across these divides, from the northern Yellowstone glacier, glacial flow was toward the other glacial termini, generally tens of kilometers away. The ice divides were generally offset toward the northern Yellowstone glacier from the present drainage divides. Thus, some ice flowed outward from the Yellowstone drainage into adjacent basins where ice levels were lower.

Obsidian hydration methods date Pinedale terminal moraines near West Yellowstone as about 30,000 years old. Pinedale glaciers probably remained near full-glacial size until 15,000 to 20,000 years ago, and deglaciation of the Pinedale icecap on the Yellowstone plateau was largely completed by 13,000 to 14,000 radiocarbon years ago.

During Pinedale recession, ice from different sources did not diminish at the same rate. The northern Yellowstone outlet glacier was still near its maximum volume when tributary mountain-valley glaciers had receded and were contributing little or no ice to it. Lakes formed where the outlet glacier blocked these tributary valleys. In Tom Miner Basin the outlet glacier advanced upvalley and deposited moraines and erratics within a few kilometers of the local cirques. Elsewhere it transported erratics at high levels across the mouths of valleys, such as Mol Heron and Bear Creeks, defining boundaries across which local ice did not advance.

At a later phase in the Pinedale recession, here termed the Deckard Flats readjustment, glaciers from the Gallatin Range and Beartooth uplift advanced into terrain previously occupied by the icecap on the Yellowstone plateau. Crossed striations and distribution of erratics indicate that their direction of flow was at angles of 60°–130° to the flow under full-glacial conditions.

A late Pinedale readvance of mountain-valley glaciers in the Gallatin Range left as many as three moraines 0.5–8 km from the cirques. At the same time, the icecap on the Beartooth uplift paused in its recession and left extensive moraines near Junction Butte.

Along the Yellowstone River in the Gardiner area, longitudinal and midchannel bars and flood-ripped fronts of alluvial fans indicate the occurrence of at least two separate floods, with waters 60 m and 45 m deep. Two additional, younger floods down the valley of the Yellowstone River resulted from the breach of a landslide dam in Yankee Jim Canyon and from release of a lake in the Grand Canyon of the Yellowstone.

Calculation of glaciological parameters, especially basal shear stress and mass balance, provides independent means for evaluating the glacial reconstruction based on field studies. Basal

shear stress was computed using the formula $\tau_b = F\rho g h \sin\alpha$. Because this formula ignores the effect of longitudinal stress gradients, its calculated value is called primary basal shear stress. Calculations were made for 50 contiguous reaches of lengths of 5–15 km along the main flow paths of the northern Yellowstone glacier and of several other glaciers heading on the same ice divide. The calculated values are all between 0.6 and 1.5 bars and are consistent with the range of 0.5–1.5 bars observed for modern glaciers. Much of the variation in primary basal shear stress is related to differing flow patterns. Primary basal shear stress averaged 1.21±0.12 bars for reaches with strongly extending flow lines, 1.04±0.16 bars for reaches with nearly uniform flow, and 0.84±0.14 bars for reaches with strongly compressing flow. These calculated values probably do not indicate actual differences in shear stress at the base of the glacier, but do suggest that the variation in primary basal shear stress is strongly related to flow pattern, especially convergence in plan view, and to ablation.

Basal shear stress calculations can also be used as an independent parameter in evaluation of alternate glacial reconstructions. In particular, an alternate reconstruction published for Yellowstone shows Pinedale ice levels 430 to 730 m lower than shown on plate 1*A* in the Washburn Range, but at the same altitudes towards the ice margin. This reconstruction results in basal shear stress values between 0.05 and 0.15 bars, as much as an order of magnitude lower than that observed for modern glaciers. Thus, in addition to being incompatible with the field observations reported in this paper, this alternate reconstruction appears glaciologically unreasonable.

The mass balance of the reconstructed northern Yellowstone glacier is calculated using (1) the area-altitude distribution shown on plate 1*A*, (2) an equilibrium-line altitude of 2,835 m as approximated by the glaciation limit, and (3) curves of specific net balance from modern glaciers. These values result in an annual net balance in which both net accumulation and ablation are equal to about 3 km^3. Because the calculated values of accumulation and ablation are approximately equal, this glacial reconstruction can meet mass-balance requirements.

The accumulation-area ratio (AAR) of the reconstructed northern Yellowstone glacier is 0.75. Although this value is high, the mass-balance calculations show that because of its area-altitude distribution, a high AAR is required for the northern Yellowstone glacier.

The discharge of the outlet glacier was approximated by three flow models. The discharge required by mass-balance data suggests that for the outlet glacier, flow by bed-slip was much greater than flowage within the glacier.

The mass-balance data when combined with very rapid rates of recession, or meltback, indicate that the increase in discharge of the Yellowstone River owing to meltback would be less than 9 percent of the discharge during the Pinedale glacial maximum. The paucity of both outwash and recessional moraines upvalley from the northern Yellowstone terminal moraines also indicates that extensive recession did not result in outwash deposition due to greatly increased discharge and sediment load.

General models, devised to show the relations in space and time among the various source areas of the northern Yellowstone outlet glacier, demonstrate the differing size and efficiency of these source areas for generating ice under full- and waning-glacial conditions (figs. 50, 51). The waxing and waning of each source area did not occur in unison, and field relations documenting incursions of ice from some sources into areas previously occupied by ice from other sources demonstrate this lack of unison. This lack of simultaneous phase relations for the growth and recession of northern Yellowstone ice masses implies that end moraines deposited elsewhere—for example, by large icecaps and small mountain-valley glaciers—are not of the same age.

INTRODUCTION

Late Pleistocene glaciation of northern Yellowstone National Park (figs. 1, 3) had several salient aspects. First, glaciers dating from the last major glaciation (Pinedale) were much larger than previously thought (pl. 1*A*). Most of the ice was 700–1,200 m thick and one continuous glacial-flow line, 146 km long, was probably the longest glacier in the conterminous United States, except for those originating in Canadian icecaps.

Second, the relative size of glaciers from several adjacent source areas changed with time, resulting in changes in flow direction of as much as 130°. Striations and directions of erratic transport, for example, show that ice in the area of the Washburn Range moved northward during full-glacial conditions but southward during deglaciation. Relations among glaciers from different source areas can be approximated; for example, during deglaciation the Yellowstone outlet glacier still retained at least 90 percent of its maximum volume at a time when local mountain glaciers had lost more than half their volume.

Third, multiple floods accompanied late Pleistocene glaciation in this area. Glacial-outburst floods probably helped build the extensive obsidian-sand plain along the Madison River near West Yellowstone. During deglaciation, multiple floods rushed down the Yellowstone valley when glacially dammed lakes released suddenly; the largest flood had a depth of about 60 m.

Fourth, the ice-distribution patterns were significantly different between Bull Lake and Pinedale time. Bull Lake moraines are far outside those of Pinedale age west of the park, but to the north they were overridden by Pinedale ice. Emplacement of large rhyolite lava flows between Bull Lake and Pinedale time blocked westward movement of Pinedale ice and resulted in larger glacial flow to the north in Pinedale time than during Bull Lake time.

The main body of this report—the Bull Lake and Pinedale glacial histories—is organized in narrative form. Factual material not essential to reconstructing the glacial history is not included here but can be found in map-reports on northern Yellowstone National Park. (See fig. 2 for index of reports.) These map-reports show the distribution of more than 40 surficial-geologic deposits at a scale of 1:62,500, describe the physical characteristics of the deposits, and give stratigraphic sections, soil profiles, flow features, and location and kind of erratics.

Finally, the high relief in northern Yellowstone made it possible to define and contour the altitude of the ice

FIGURE 1.—Yellowstone National Park region. Hachured lines, outer limits of end moraines of probable Pinedale age; line labeled with B's, outer margin of probable Bull Lake moraines. Moraine positions in the West Yellowstone Basin are modified from Alden (1953), Richmond (1964a), Pierce (1973a), Waldrop and Pierce (1975), and Waldrop (1975a). Moraines north of the park along the Yellowstone valley are modified from Weed (1893) and Montagne (1970); also see this report (pl. 1*A*; fig. 12). Moraines along both forks of Rosebud Creek and Rock Creek are from Ritter (1967). Moraines along the South Fork of Shoshone River are from Pierce and Nelson (1969). Moraines in Jackson Hole are from Love and Reed (1971). Moraines southwest of the park are from Richmond (1973a). Shaded-relief base map made from photograph of Army Map Service pressed-relief maps.

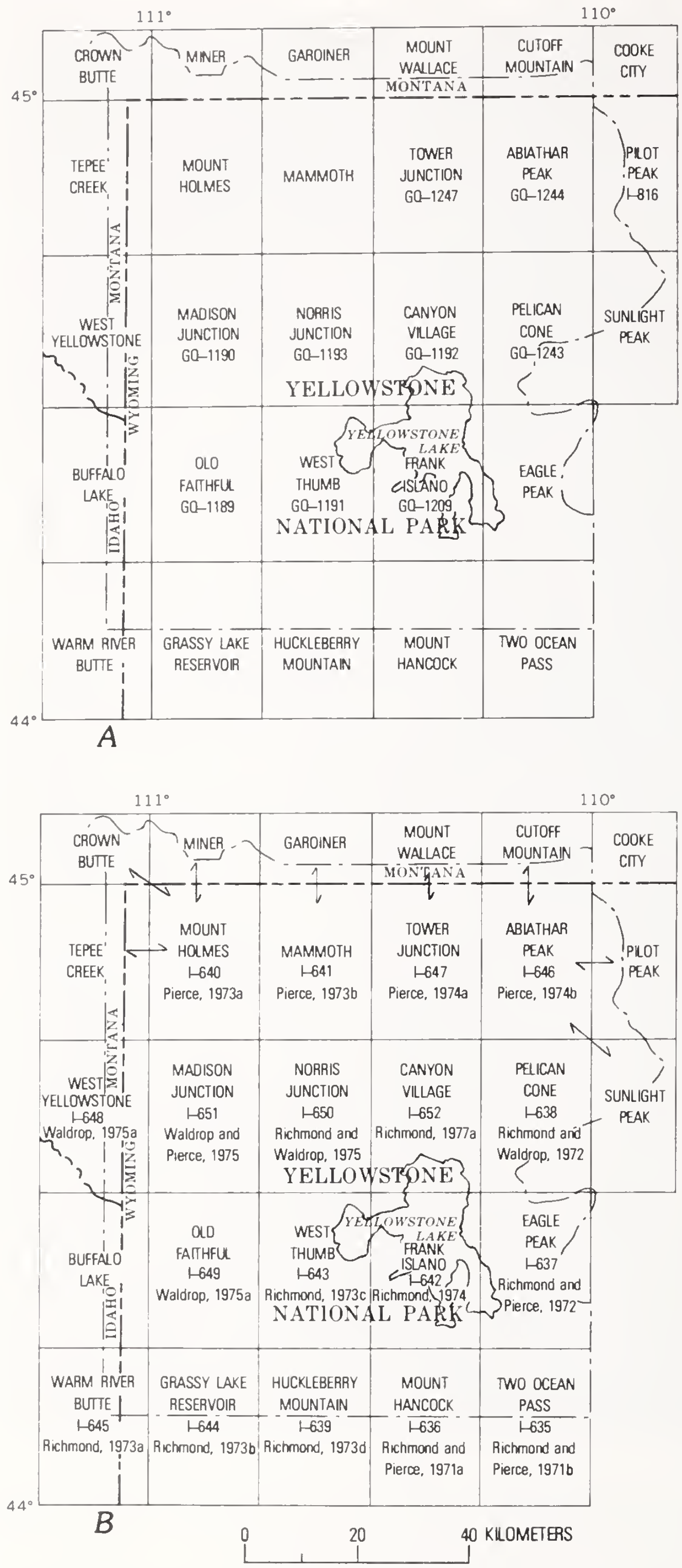


FIGURE 2.—Locations of surficial and bedrock maps of Yellowstone National Park, at a scale of 1:62,500. *A*, Bedrock geology (also shown at a scale of 1:125,000 (U.S. Geol. Survey, 1972b)). *B*, Surficial geology (also shown at a scale of 1:125,000 (U.S. Geol. Survey, 1972a)).

surface. Contours on the ice surface are the key parameter needed in both the basal shear stress and mass-balance calculations.

PREVIOUS WORK

Magazine articles published by Geikie (1881) and Holmes (1881) stated that the Yellowstone Park area had been extensively glaciated. Geikie concluded that a glacier thousands of feet thick had flowed northward from the park down the Yellowstone valley. Holmes (1881; 1883) determined, mainly on the basis of glacial erratics, that ice had covered much of the Yellowstone plateau. Weed (1893) provided the first detailed map and description of glaciation in the Yellowstone valley north of the park; his reconstruction of the northern Yellowstone glacier north of the park is essentially the same as that presented here for Pinedale full-glacial conditions. Howard (1937) made an award-winning study of the glacial features and history of the Grand Canyon of the Yellowstone. The terminal moraine area of the Yellowstone valley was studied by Horberg (1940). Miner (1937) wrote a thesis on the Mammoth area, dealing mainly with features of the Pinedale deglaciation. Alden (1928), in a naturalist's guide for interpreting Yellowstone to the public, appears to have clearly understood the main elements of the last two glaciations of Yellowstone, and later (1953, p. 177–179) he formally published a study of the glacial geology of the West Yellowstone area. Hall (1960, 1961) studied the glaciation of the area just northwest of the park. Love (1961) described widespread faulting of Quaternary deposits in Yellowstone. John Montagne worked extensively in the Yellowstone valley and in the Gallatin Range (1968, 1970); although much of his work is not yet published, I was able to benefit from his knowledge through numerous field trips and discussions. As one result of these discussions, we concluded in 1970 that the Chico and Eightmile moraines in the Yellowstone valley were both of Pinedale age.

Following the Hebgen Lake earthquake of 1959, the U.S. Geological Survey made a study of West Yellowstone area: Richmond (1964a) mapped and described the surficial geology of the area and established the precedent for the Bull Lake–Pinedale nomenclature used in the ensuing U.S. Geological Survey reports. Richmond and Hamilton (1960) recognized that the rhyolite flow now called the West Yellowstone flow is younger than the nearby Bull Lake moraines.

As part of the present U.S. Geological Survey study of Yellowstone Park, surficial geologic maps at a scale of 1:62,500 with accompanying texts have been prepared by G. M. Richmond, H. A. Waldrop, and K. L. Pierce; these map-reports are indexed in figure 2.

ACKNOWLEDGMENTS

The comprehensive geological study of Yellowstone National Park, of which this report is a chapter, was inaugurated by the late Arthur B. Campbell, U.S. Geological Survey. It was a combined research project sponsored by NASA that involved studies of the Yellowstone geology as well as an evaluation of various remote-sensing techniques. This study was immeasurably aided by the full cooperation of the Park Service. John Good, Chief Naturalist of Yellowstone National Park, and his successor William Dunmire, made the local arrangements necessary for this extensive survey.

G. M. Richmond served as project chief for studies of the surficial geology of Yellowstone; his wide experience in Rocky Mountain glacial geology provided a framework for this study. The study has benefited greatly by discussions with D. R. Crandell, R. F. Madole, S. M. Colman, H. A. Waldrop, W. G. Pierce, John Montagne, A. R. Southard, F. L. Nials, W. B. Hall, Irving Friedman, W. E. Scott, and H. E. Walsh. M. F. Meier made some of the flow calculations and provided other glaciological advice. R. M. Burke kindly provided unpublished analytical data on the soils in the West Yellowstone Basin. I was ably assisted in the back country field work by R. B. McIntyre in 1965, Clifford Montagne in 1967 and 1969, and D. H. Lehman in 1968.

PHYSICAL FEATURES RELEVANT TO GLACIATION

REGIONAL TOPOGRAPHY

The northern Yellowstone area consists of high mountains, plateaus, and deep valleys (figs. 1, 3).

The high mountains north and northeast of the park are here referred to as the Beartooth uplift (fig. 3; pl. 1*A*; fig. 5), which is a large, relatively coherent structural and geographic block. The eastern part of the uplift forms the Beartooth Mountains. The western part previously has been called the Snowy Range (Hague, 1912; Alden, 1928; Howard, 1937), the Beartooth Mountains (Foose and others, 1961), and is now officially designated as part of the Absaroka Range.

The term Absaroka Range is restricted in this report to the mountains so named south from the Cooke City, Mont., area to the southern boundary of the northern Yellowstone National Park area (fig. 3; pl. 1*A*; fig. 48). It does not include mountains of the Beartooth uplift. The Gallatin Range extends, and widens, northward from the northwestern part of the park.

The Washburn Range, Specimen Ridge, and Mirror Plateau form uplands in the north-central part of the park; the rest of the area, here called the Yellowstone plateau, is mostly an undulating upland at an elevation of 2,400±200 m (8,000±600 ft).

All of northern Yellowstone National Park drains eastward toward the Mississippi River; the Continental Divide passes through the southern part of the park. The Yellowstone River and its tributaries drain the eastern three-quarters of the park. The Yellowstone River rises southeast of the park, flows out of the north end of Yellowstone Lake, across Hayden Valley, and then through its Grand Canyon and Black Canyon. The eastern boundary of the park is a high divide that separates the Yellowstone drainage from the Clark's Fork of the Yellowstone drainage and the Shoshone River drainage. The western quarter of northern Yellowstone is drained by the Madison and Gallatin Rivers.

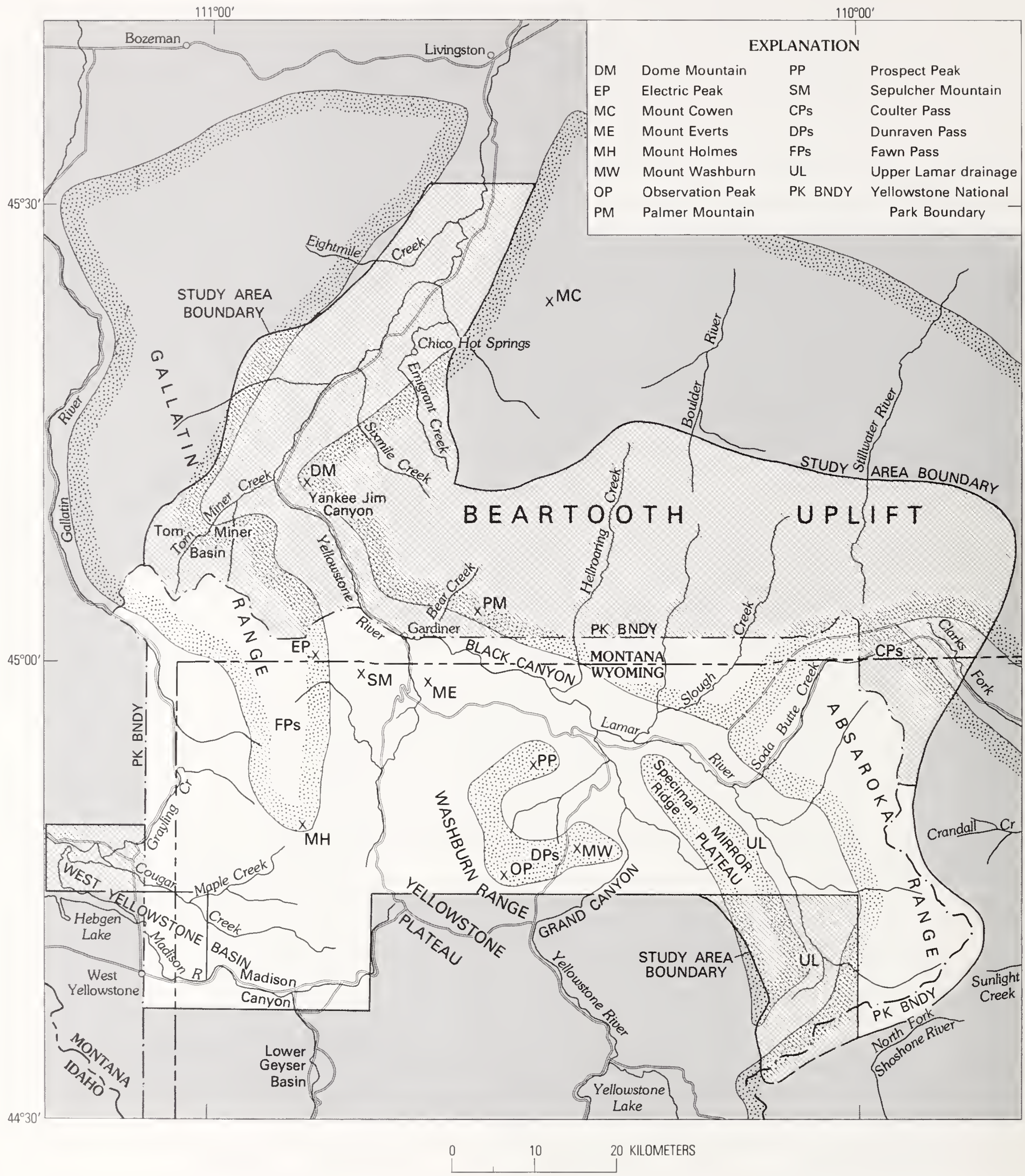
BEDROCK TYPE AND DISTRIBUTION

A great many bedrock types occur in the northern Yellowstone area (pl. 2). Earlier studies include those of Hague and others (1899); Hague (1904; 1912); Howard (1937); Seager (1944); Brown (1961); Boyd (1961); Hall (1961); Fraser, Waldrop, and Hyden (1969); and Witkind (1969). Recent mapping by U.S. Geological Survey geologists is referenced in figure 2; recent bedrock reports include those on the Gallatin Range by Ruppel (1973), on the Quaternary volcanics by Christiansen and Blank (1972), on the Eocene volcanics by Smedes and Prostka (1972), and on the Beartooth uplift by Wedow and others (1975).

The bedrock types differ markedly in the degree to which they preserve evidence of glaciation. For example, Precambrian granitic rocks form and clearly retain evidence of glacial erosion, but they unfortunately are not very common in northern Yellowstone National Park. The two most widespread rock types—Quaternary rhyolite (both flows and welded tuffs) and Eocene andesitic volcanic rocks of the Absaroka Volcanic Supergroup (breccias, conglomerates, sandstones and mudstones)—do not normally preserve evidence of glacial erosion. Moreover, till derived from these rocks is difficult to identify as such. These are probably the main reasons the magnitude of Pinedale Glaciation of northern Yellowstone has not been recognized previously.

CLIMATE

The climate of Yellowstone is severe. Mean annual temperatures on the Yellowstone plateau are less than 1°C, and snowfall is deep (see section "Specific net balance estimates") and covers the ground about half the year. Figure 4 shows the temperature, snowfall,


111°00′
110°00′
45°30′
45°00′
44°30′
EXPLANATION
DM Dome Mountain
EP Electric Peak
MC Mount Cowen
ME Mount Everts
MH Mount Holmes
MW Mount Washburn
OP Observation Peak
PM Palmer Mountain
PP Prospect Peak
SM Sepulcher Mountain
CPs Coulter Pass
DPs Dunraven Pass
FPs Fawn Pass
UL Upper Lamar drainage
PK BNDY Yellowstone National Park Boundary
Bozeman
Livingston
Eightmile Creek
STUDY AREA BOUNDARY
MC
Chico Hot Springs
Emigrant Creek
Sixmile Creek
Boulder River
Stillwater River
STUDY AREA BOUNDARY
GALLATIN
RANGE
Gallatin River
River
DM
Yankee Jim Canyon
Tom Miner Creek
Tom Miner Basin
BEARTOOTH UPLIFT
Hellroaring Creek
Bear Creek
PM
Yellowstone River
Gardiner
PK BNDY
Slough Creek
Clarks Fork
EP
BLACK CANYON
MONTANA
WYOMING
CPs
SM
ME
FPs
Lamar River
Soda Butte Creek
ABSAROKA RANGE
PK BNDY
PP
Speciman Ridge
MIRROR PLATEAU
UL
Grayling Cr
MH
WASHBURN RANGE
DPs
MW
OP
Crandall Cr
Cougar Creek
Maple Creek
WEST YELLOWSTONE BASIN
Hebgen Lake
Madison R
YELLOWSTONE PLATEAU
GRAND CANYON
STUDY AREA BOUNDARY
UL
West Yellowstone
Madison Canyon
Yellowstone River
PK BNDY
Sunlight Creek
North Fork Shoshone River
MONTANA
IDAHO
Lower Geyser Basin
Yellowstone Lake
0 10 20 KILOMETERS

and precipitation for stations in and near Yellowstone.

Precipitation is strongly related to altitude. Also, for a given altitude, precipitation increases toward the southwest part of the park. This increase results from the rise from the Snake River Plain to the Yellowstone plateau of Maritime Polar and Maritime Pacific air-masses (Dirks, 1974, p. 32). The prevailing winds are from the southwest.

All the weather stations are below an altitude of 2,400 m (fig. 4); consequently, estimates of the precipitation on the higher parts of the Yellowstone plateau and on all the mountain ranges must be based on study of snow-survey and streamflow records, and on vegetation associations. For the mountain ranges in northern Yellowstone, isohyets drawn by P. E. Farnes (in Dirks, 1974, fig. 3) show that precipitation at higher altitudes commonly exceeds 100 cm. The following are representative values for the higher altitudes in the different ranges of northern Yellowstone: Gallatin Range at 2,900 m (9,500 ft), 150 cm; Washburn Range at 2,700 m (8,850 ft), 100 cm; Specimen Ridge at 2,800 m (9,200 ft), 130 cm; and Beartooth uplift at 3,400 m (11,150 ft), 180 cm.

The climate is cold enough to preserve or form permanent ice (permafrost) in the ground at favorable sites. Local permafrost occurs as low as 2,100 m (6,800 ft) on a north slope near Tower Junction (Good, 1964), at 2,600 m (8,500 ft) in Sylvan Pass, and at 2,400 m (7,800 ft) in the Gallatin Valley.

EFFECT OF GEOTHERMAL FEATURES ON GLACIATION

The heat produced by Yellowstone's famous geothermal features affected glaciation. The hot springs and geysers of the Lower, Midway, and Upper Geyser Basins produce enough heat annually to melt ice 5 m thick over the entire area of the geyser basins, or 0.3 m thick if averaged over the entire Firehole River drainage (Fournier and others, 1970). In comparison to melting at a glacier's surface, 0.3 m of basal melting is about equivalent to the increase in surface melting that accompanies a 30-m decrease in altitude in a glacier's ablation area (see fig. 49). The thermal areas evidently were unable to melt ice as fast as it was supplied, for all the major thermal areas were covered by ice during the last glaciation (U.S. Geological Survey, 1972a). But when glacial recession resulted in the ice margin being just beyond thermal areas, geothermal heat commonly melted holes or embayments into the ice in which lake and stream deposits accumulated. The Upper, Midway, and Lower Geyser Basins are ideally located to melt an embayment into the western margin of the Yellowstone icemass, forming a lake behind an ice dam near Madison Junction (Waldrop and Pierce, 1975). The sudden release of such lakes may be responsible for the obsidian-sand plain near West Yellowstone.

Normal heat flow of 1.5×10^{-6} cal/cm^2/sec is capable of annually melting a thickness of about 7 mm of ice from the bottom of a glacier. Heat flow inside the Yellowstone caldera (Christiansen and Blank, 1972) may be several times this amount and hence capable of melting several centimeters of ice per year.

◁ FIGURE 3.—Northern Yellowstone area, including source areas and terminus of the large icemass that covered almost all of northern Yellowstone National Park. Unshaded area was mapped in detail at a scale of 1:62,500 (Pierce, 1973a, b; 1974a, b; Waldrop and Pierce, 1975). Diagonal line pattern area includes the following supplemental areas containing parts of the northern Yellowstone icemass: (1) The glacial geology along the Yellowstone River valley north of the Park is based on studies by Weed (1893), Horberg (1940), Montagne (1968, 1970, and written commun., 1972), and my own reconnaissance and local detailed studies. (2) That for the Beartooth uplift is based on air-photo interpretation of more than 450 sites of glacial-scour features and reconnaissance field studies by me, and on discussions with J. E. Elliott, F. S. Simons, D. L. Gaskill, W. G. Pierce, and H. J. Prostka, all of whom have recently mapped bedrock geology in this area. (3) That south and west of the Madison Canyon is based on studies of Waldrop (1975a,b) and Waldrop and Pierce (1975). (4) That northwest of West Yellowstone, is based on studies by Alden (1953), Richmond (1964a), and myself. (5) And that in the southwestern part of the upper Lamar drainage is based on studies of Richmond and Waldrop (1972).

PATTERNS OF ICE ACCUMULATION, MOVEMENT, AND ABLATION

The glaciation of northern Yellowstone cannot be understood unless one considers the several different sources and their interrelations through time. Under full-glacial conditions (when glaciers were at or near their maximum sizes), ice bodies from various sources merged and the position of their shared boundary was defined by the combined flow pattern. During recession, ice from some sources receded more rapidly than that from others, commonly resulting in movement of ice from the more persistent sources into places previously occupied by ice from the less persistent sources.

Although size and elevation of source areas were probably the most important factors influencing the productivity of glacial-source areas, geographic location with respect to incoming snowstorms was also very significant. Many storms probably moved eastward up the Snake River Plain and dropped great amounts of snow when the air masses rose over the icecaps on the Yellowstone plateau.

The following paragraphs describe the geometry of the various sources and flow outlets and their

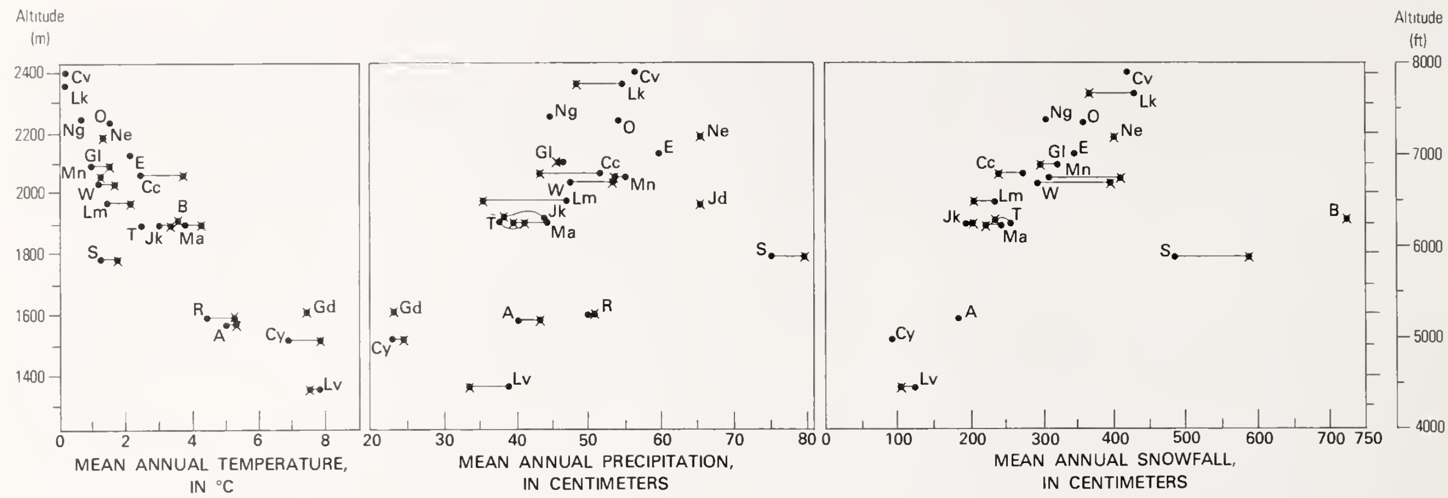


FIGURE 4.—The relation of mean annual temperature, precipitation, and snowfall to altitude in the Yellowstone region. Dot indicates values recorded before 1930, X after 1930 (from U.S. Weather Bureau, 1934, 1960). Snowfall is the total of individual snowfalls, not the total depth of snow on the ground. The stations are in Wyoming unless otherwise noted.

A—Ashton, Idaho
B—Bechler Ranger Station
Cv—Canyon Village (originally recorded as Grand Canyon)
Cc—Crandall Creek
Cy—Cody
E—East Entrance
Gd—Gardiner, Montana
Gl—Gallatin Ranger Station, Montana
Jd—Jardine, Montana
Jk—Jackson
Lk—Lake Ranger Station
Lm—Lamar Ranger Station (originally recorded as Buffalo Ranch)
Lv—Livingston, Montana
Ma—Mammoth (commonly recorded as Yellowstone National Park)
Mn—Moran (in Jackson Hole)
Ne—Northeast Entrance
Ng—Norris Geyser Basin
O—Old Faithful (originally recorded as Upper Geyser Basin)
R—Red Lodge, Montana
S—South Entrance (originally recorded as Snake River)
T—Tower Falls
W—West Yellowstone (includes Riverside), Montana

characteristics through waxing-, full-, and waning-glacial conditions. The final section of this report contains graphical models (figs. 50, 51) of the ice sources in both space and time. Although these figures represent conclusions, the reader will find them helpful in understanding the attributes of the major sources.

MAJOR ICE SOURCES AND DRAINAGEWAYS

The major ice sources and drainageways are diagrammatically shown in figure 5. The sources responded differently to the glacial climatic cycle. These differences in response are a major theme of this report.

The major source areas supported icecaps. About half of the flow from these icecaps converged to form the northern Yellowstone outlet glacier. The boundary separating flow toward the northern Yellowstone outlet glacier from flow away from it was generally near, but seldom coincident with the present drainage divides.

INTERACTION BETWEEN GLACIERS FROM DIFFERENT SOURCES THROUGH TIME

The volume of ice from the different source areas did not increase or decrease in the same relative proportions during changes in climate. Small mountain-valley glaciers seem to have responded much more quickly to climatic change than did glaciers from large source areas such as the Beartooth uplift; this effect is analogous to inertia. The icecap on the Yellowstone plateau illustrates a threshold-type effect that was self-amplifying as full-glacial conditions were approached. Low- to intermediate-glacial conditions were probably not sufficient to *initiate* an icecap on the Yellowstone plateau, but they probably were sufficient to maintain one; during deglaciation the plateau icecap rapidly stagnated.

Other second order, time-related effects were also probably involved. As the icecap on the Yellowstone plateau thickened and expanded, it caused more and more orographic snowfall on its surface, and thus reduced snowfall in the Absaroka Range and Beartooth uplift downwind from the icecap.

In describing the change in ice flow and ice occupation of an area, this report will use a domain concept: domain is referenced to the areas occupied by ice from various sources under full-glacial conditions. If ice from one source moves into what was the full-glacial domain of another source area this is an "incursion." By use of the domain concept flow directions and relative ice-productivity of different source areas can thus be related through time.

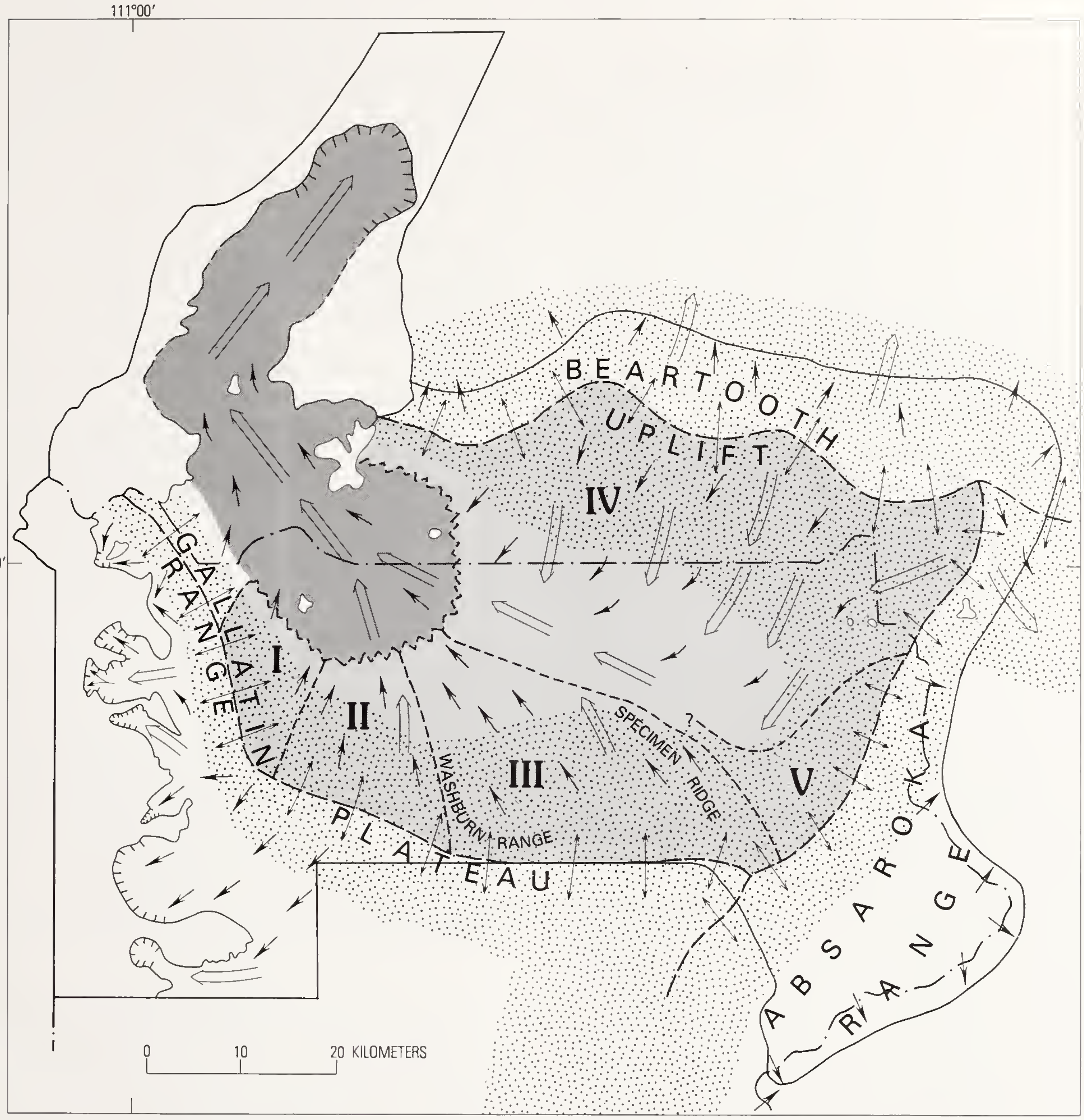


FIGURE 5.—Major full-glacial ice-sources and drainageways of northern Yellowstone National Park. Light and dark shading, accumulation and ablation areas of northern Yellowstone galcier. This glacier (heavier shading) terminated in the Yellowstone valley and was fed by parts of interconnected icecaps from five main source areas (open stipple). The short-dashed lines separate the source area of this glacier into five flow components that head on: (I) the Gallatin Range, (II) the plateau between the Gallatin and Washburn Ranges, (III) the plateau south of the Washburn Range, (IV) the Beartooth uplift, and (V) the upper Lamar drainage. Double-point arrows on heavy dashed lines represent flow away from ice divides. Broad, open arrows, major ice drainageways more than 800 m deep; small arrows, flow directions; hachured lines, Pinedale terminal moraines in the northern Yellowstone area. Outline of ablation area dashed where uncertain.

DISCUSSION OF METHODS

Most methods employed in this study are standard and are described more completely in texts such as Flint (1971). The methods discussed in detail here include only those which are either not widely known or that require reevaluation with regard to their usefulness in this study. Many others are briefly described where relevant in the section on glacial history.

INDICATORS OF AGE

The following criteria were used to determine age relations of glacial features in northern Yellowstone:

- I. Indicators of the relative magnitude of age difference
  - A. Weathering criteria
    1. Extent of soil development
    2. Thickness of weathered rinds on stones
    3. Degree of muting of morphologic expression
    4. Weathering and pitting of rock surface
    5. Presence and thickness of loess mantle
  - B. Radiometric age determinations (K-Ar, ^{14}C)
  - C. Obsidian-hydration dating
  - D. Geometric relations to deposits of known age
  - E. Amount of offset along the same fault
  - F. Relation to volcanic ash beds
  - G. Pollen profiles from overlying lacustrine or bog sediments
- II. Indicators of relative age which are independent of the magnitude of age difference
  - A. Cross-cutting relations in plan view
    1. Moraines
    2. Cross-cutting striations
    3. Cross-cutting scour features
    4. Ice dams, backfills, changes in erratic-transport directions
  - B. Sequence relations
    1. Position relative to ice source
    2. Stratigraphic sequence
    3. Age relations required by reconstruction of geologic history

SOIL DEVELOPMENT

The degree of development of relict soil profiles has been used in the Rocky Mountains to group glacial deposits by age (Richmond, 1962a). A major effort was expanded in northern Yellowstone to do the same. Unfortunately, the great variability of parent materials there complicates matters to the extent that soils could not be used as an unambiguous age-criterion.

The degree of soil-profile development is most readily estimated by the amount of clay enrichment, the expression of soil structure, and the thickness of the B-horizon. Soil development is generally aided in northern Yellowstone by a wet climate and a significant amount of clay in the parent material.

Relict soils on Bull Lake deposits generally show stronger development than those on Pinedale deposits, although the amount of variability in development of soils on Pinedale deposits overlaps that on Bull Lake deposits. Before ages can be assigned to soil profiles, one must account for the effect of factors such as parent material, completeness of preservation of the profile, susceptibility to illuviation, topographic effects, and climate on soil development. Thus, although degree of soil development is useful as an age criterion, like most other Quaternary relative-age criteria, it necessitates subjective judgments.

Variation in parent material alone complicates the use of relict soils in northern Yellowstone. For example, post-Pinedale soils developed on sandy till derived from rhyolite are poorly developed compared to those on till containing moderate amounts of clay.

Moreover, it is generally not possible to distinguish primary from inherited characteristics. Some soils start with what might be called a "stacked deck," in terms of their ability to exhibit a significant amount of soil "development" in a short time. In some cases, this may be literally a "stacked deck" consisting of a fine-grained layer containing easily mobilized clay overlying a moderately porous layer: the clay may be readily washed down into the underlying deposit, forming a B-horizon enriched in clay. However, such a B-horizon is not made up of clay-minerals formed by in-place weathering. A similar condition is produced when fine-grained airborne silt and clay are deposited on the surface and washed down into the soil profile.

Some clays, apparently of the swelling type, readily form well-developed soil profiles, when mixed with coarser material in the approximate ratio of one-third clay to two-thirds coarser material. Some clays derived from Paleozoic or Mesozoic shales or Eocene mudstone seem to migrate readily downward in soil profiles, perhaps as a result of expansion and contraction of the clays during wetting and drying cycles. Such cycles are characteristic of northern Yellowstone, where soils deeply saturated by snowmelt are dry by late summer. The clay moves from the upper part of the soil into the B-horizon where it forms a textural B-horizon. In this instance, the amount of clay in the B-horizon is not a function of mineral-grain weathering, but merely the movement of clay. Wetting and drying of a clayey B-horizon also promotes development of soil structure in a relatively short time.

OBSIDIAN HYDRATION

The thickness of hydration rinds has been used to date obsidian artifacts and rhyolite flows in the time interval from several hundred to several hundred thousand years ago (Friedman and Smith, 1960; Friedman, 1968). A fresh surface of obsidian (nonhydrated volcanic glass of rhyolitic composition) will absorb water from the air or soil, and this water will slowly diffuse into the obsidian at a rate that is independent of humidity, but dependent mainly on temperature and the composition of the obsidian. The diffusion front is sharp and can be recognized by the abrupt change in refractive index at the hydration interface. The hydration rate relative to temperature has been experimentally determined (Friedman and Long, 1976).

Experimental evidence and data from archeological material indicate that the thickness of the hydration rind increases linearly with the square root of time:

$$\text{thickness} = \text{constant} \times (\text{time})^{1/2} \text{ or}$$
$$(\text{thickness})^2 \propto \text{time}.$$

Hydration will start whenever a fresh surface of obsidian is exposed, such as when an obsidian pebble is abraded, a rhyolite flow cracks during cooling, or an obsidian artifact is manufactured. Experimental determination of hydration rates with variable water pressure, and inspection of hydration rinds on obsidian from archeological sites in "wet" and "dry" environments both show that relative humidity is unimportant in the hydration process (Friedman and Smith, 1960). In order to minimize annual or diurnal temperature changes, samples for dating should be collected from at least 1.5 m below the natural ground surface. Thermal areas are not appropriate for sampling because of abnormally high ground temperatures.

Various thicknesses of rinds are commonly noted on thin sections of obsidian pebbles from a glacial deposit. The thickest rinds develop adjacent to the original cooling cracks; they generally occur beneath smooth arcuate surfaces and around phenocrysts and vesicles. Thinner rinds date from times when an obsidian fragment was glacially abraded, and the thickness of these rinds may be used to date glacial deposits (Pierce and others, 1976).

Two forms of cracks not related to cooling are common in obsidian from glacial deposits. Scallop cracks are subparallel to the pebble surface and are thought to be produced when the pebble was forcibly dragged across another hard object. Pressure cracks occur in groups that extend from the clast surface toward like fingers; they are thought to result from intense pressure on a small area of the obsidian clast in contact with another hard object (Pierce and others, 1976, fig. 2).

DETERMINING DIRECTION OF GLACIAL FLOW

Erratics.—Glacial erratics are one of the most commonly used indicators for reconstructing directions of ice movement. In northern Yellowstone, where multiple ice sources did not necessarily function in unison and multiple glaciations are involved, erratics may have been carried in different directions at different times during a single glaciation, or they may have been carried in one direction during one glaciation, and in a different direction during subsequent glaciation. Consequently, erratics that do not make up a significant portion of a glacial deposit are used only to indicate net transport direction and are not used to indicate flow direction during any specific phase of glaciation unless supported by other evidence.

Rat tails.—Glacial pavement on rocks of varying resistance, especially conglomerates containing hard pebbles, commonly display rat tails. These are streamlined forms on the down-ice side of resistant clasts that stand higher than the surrounding rock (figs. 6*A*, 7*A*). Rat tails are probably the most obvious and certain indicator of the direction of glacial flow. Unfortunately, the main conglomerates present in northern Yellowstone, from the Eocene Absaroka Volcanic Supergroup, are so poorly indurated that glacial pavement is seldom preserved on them, even on cirque floors.

Striations.—Glacial striations may indicate either regional or local flow patterns. They establish the line of flow, but do not necessarily indicate which direction the ice moved. If the striations occur on a peak or ridge crest well above other nearby topographic features, they closely reflect the direction of regional flow and parallel the direction of slope of the ice surface. If they occur within a valley and parallel the valley, their orientation may be controlled by local topography. But striations and other flow features that trend across the topographic grain tend to indicate regional flow and, thus, the direction of slope of the ice surface.

Some striations can be used to determine the sequence of glacial flow directions. For example, a set of glacial striations may cut across an older set. Or striations with different orientations may occur on high and low parts of the same ridge crest; the higher striations indicate an earlier flow direction and the lower ones a later direction.

Striations and polish on bedrock protuberances.—The preferential development of polish and striations on the up-ice side of a bedrock protuberance compared to that on the down-ice side is a common indicator of the direction of flow. Clasts 0.2–1 m across in the Eocene bedrock form protuberances that commonly show such preferential development of striations and polish. This is the result of compressive flow as the base of a glacier crowds around the protuberance (figs. 6 and 7). Striations indicate the alinement of flow, and the distribution of polish and striations shows the direction of flow along this alinement.

Polish on treads and not on risers.—Ice flow in some places was across a stairstep-like set of benches. Where the risers face the direction of flow, striations and polish are developed on the outer portions of the steps and are absent on the risers (figs. 6*C*, 7*B*).

Low-angle friction cracks.—Fractures sometimes form in bedrock when it is overridden by ice and its incorporated debris. A variety of "chatter marks" identified by their shape in plan view have been described (Gilbert, 1906; Harris, 1943; Okko, 1950; Flint, 1971). In the present study, only low-angle friction cracks were used as a directional indicator (figs. 6*D*, 7*C*).

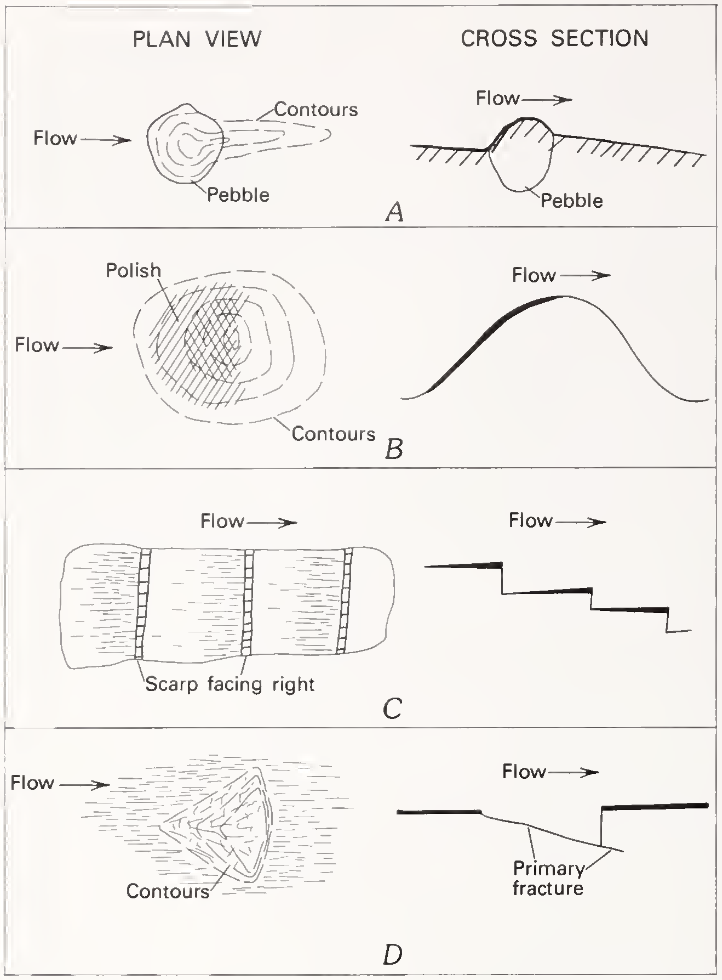


FIGURE 6.—Some small-scale features used in this study to determine direction of glacial flow. These features are particularly important in the Washburn Range–Specimen Ridge area, where they indicate full-glacial flow northward toward the outlet glacier at Gardiner. *A*, Rat tail; *B*, Striations and polish on upflow side of protuberance; *C*, Striations and polish on tread and not on riser; *D*, Low-angle friction crack. Heavier lines in cross sections indicate places of glacial polish and striations.

These cracks dip in the direction of ice-flowage. They seem to be analogous to tension gashes found along fault surfaces, with the contact between the debris-laden glacier and bedrock being analogous to the fault surface. Flint (1971, p. 97; 1955, p. 59) and Dreimanis (1953) concluded that not all friction cracks dip in the direction of flow and cited examples of cracks inclined in the opposite direction at angles of 50°–90° to the rock surface. Cracks inclined at an angle of more than 35° to the rock surface were not used as flow-direction indicators in this study so as to eliminate the problem of high-angle friction cracks.

RECONSTRUCTION OF THE ICE SURFACE

A reconstruction of the shape of the glacier surface in northern Yellowstone was a major objective of this study; such a reconstruction is possible in detail only for the last (Pinedale) glaciation (pl. 1). Areas covered by ice during the last glaciation are defined by end moraines, lateral moraines, erratics on glacially scoured terrain, large-scale glacial-scour features (apparent on aerial photographs and on the ground), and fresh glacial striations and polish. In transects that extend from glaciated to unglaciated terrain, it is commonly possible to determine the altitude of the ice surface within 30 meters.

Under equilibrium conditions in temperate latitudes, about two-thirds of the surface area of an ice mass is in the glacier accumulation area above snowline and about one-third is in the ablation area below snowline. Glacial deposition near the ice margin occurs only in the ablation area. Thus, in determining the maximum height of a glaciation, only erosional features are to be expected over most of a former glacial system, with the exception of recessional glacial deposits along valley floors in the source area.

The best places for determining ice-surface altitudes are provided by nunataks. The minimum height of ice of the last glaciation is indicated by the highest fresh evidence of glaciation. The maximum height of ice is more difficult to determine accurately: it is approximated by the lowest altitude where fresh evidence of glaciation was not observed, but should have been preserved had the area been glaciated. If all the high places in a given area were covered, as was commonly the case in northern Yellowstone, the glaciated high points only indicate the minimum altitude of the ice surface.

The orientation of contours on the ice surface and the direction of surface slope can be determined if appropriate glacial scour or other features are observed. With these form lines, the surface altitude of the ice can then be projected laterally from places where the actual altitude of the ice has been determined from field observations on nunataks, or where a minimum altitude has been determined on glaciated high points. One needs to be able to infer that glacial scour features relate to full-glacial flow and not to some anomalous local topographic or recessional effect. Striations on ridge crests near the upper limit of glaciation provide the best indicators of the direction of full-glacial flow.

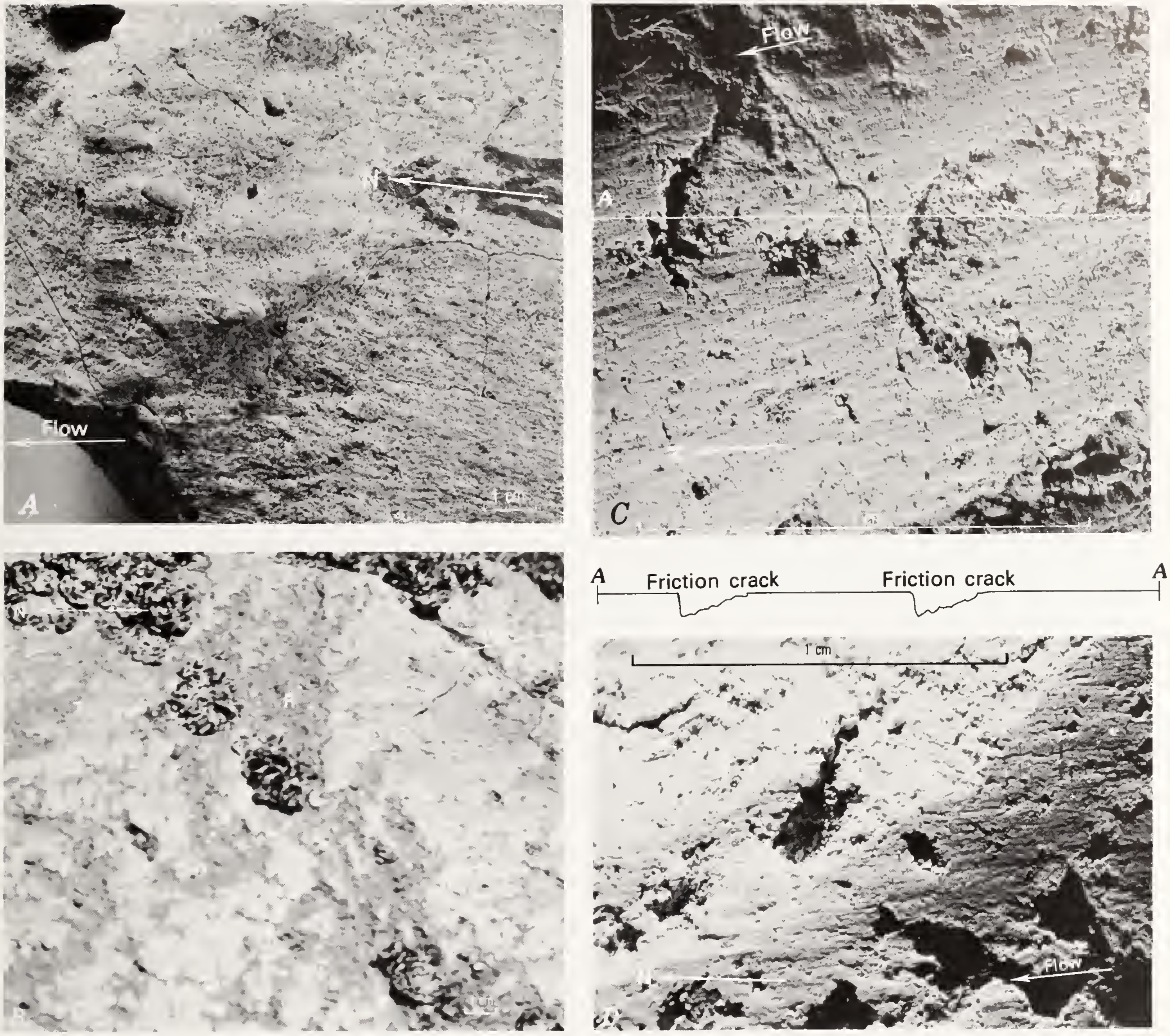


FIGURE 7.—Directional glacial-scour features. *A*, Rat tails on downflow sides of conglomerate pebbles. Flow to the left parallel to striations. Locality on north side of Mount Washburn at altitude of 2,560 m (8,400 ft). *B*, Polish on tread, T, and not on riser, R. Riser facing to the left is not polished, indicating flow to left parallel to striations. Locality on east side of Mount Washburn at altitude of 2,620 m (8,600 ft). *C*, Photomicrograph of small friction cracks. Flow parallel to striations and toward the left. Sample collected from summit area of Mount Washburn and coated with thin white film for photographic enhancement. *D*, Photomicrograph of striations rounding a protuberance, summit area of Mount Washburn. Polish on upflow side of corner. The hackly area just left of center is part of a friction crack.

THE BULL LAKE–PINEDALE PARADIGM

Following Blackwelder's (1915) original description of end-moraine sequences in the Wind River Range, in which older, more muted Bull Lake moraines were typically succeeded upvalley by fresher Pinedale ones, many geologists have mapped similar morainal successions elsewhere in the Rocky Mountains. This Bull Lake–Pinedale succession has been so widely recognized that one normally expects to find it, and it has become a widely accepted paradigm for the subdivision of glacial deposits in the Rocky Mountains. From experience in Yellowstone and elsewhere, I think the Bull Lake–Pinedale paradigm tends to produce a similar pattern of age assignments, regardless of the actual age of the deposits.

In studies of end-moraine sequences, a common procedure is to examine a sequence for differences in boulder abundance, preservation of original morphology, soil development, and other relative-age characteristics. For the purposes of correlation with the standard Rocky Mountain sequence, one looks to the outer moraines as possible candidates for a Bull Lake age assignment, and to sequentially younger moraines as candidates for a Pinedale age assignment. The problem then becomes to determine if relative-age criteria indicate this age difference. Quantitative measures of this difference are uncommon, and are quite incomplete in the type areas.

Such a relative-age difference is not one of absolute values that are comparable from range to range or valley to valley, mainly because of differences in climate, rock type, and texture of the deposits. Thus, in the absence of numerical dates, the differences within a morainal sequence are entirely relative. Whether or not the relative-age differences are of the Bull Lake–Pinedale magnitude is a subjective decision. Also, the observations are commonly subjective; a major age criterion is the visual evaluation of moraine morphology. Examination of almost any end-moraine sequence produces some differences in relative-age criteria. Seldom are statistical measures employed to determine if the differences are meaningful. The researcher normally is motivated to find differences, because concluding that he can find no significant age differences in a moraine sequence would be failing to find what others have commonly found. Therefore, there is a bias toward looking for differences, and an expectation that the differences found might be of the magnitude of the Bull Lake–Pinedale contrast.

However, if one considers the relative size of the Bull Lake and Pinedale Glaciations as commonly mapped, an alternative hypothesis should be entertained. In many places in the northern Rocky Mountains (for example, western Mount Cowen, fig. 12), Pinedale glaciers were at least 95 percent as long as the preceding Bull Lake glaciers. Thus only a 5 percent (or less) increase in size of Pinedale glaciers would result in overrunning and obliterating Bull Lake terminal moraines. In some places where both Pinedale and Bull Lake moraines are identified, Pinedale glaciers were clearly longer than Bull Lake ones, and Bull Lake moraines are preserved only because Pinedale ice flowed in a different direction in the terminal area (for example, Strawberry Creek, fig. 12). It is likely that in some places Pinedale glaciers have overridden and obliterated Bull Lake moraines, leaving an end moraine sequence only of Pinedale age.

Other considerations also complicate the application of the Bull Lake–Pinedale paradigm. Recent studies suggest that in some areas Pinedale deposits differ in age by a factor of 2 or more (Pierce and others, 1976; S. M. Colman, written commun., 1977). Thus, appropriately sensitive relative-age criteria may show significant age differences within a succession of Pinedale age. Therefore, even if an end moraine sequence is all of Pinedale age, it is possible to find some relative age criteria indicating an age difference. Also because one normally anticipates a downvalley increase in age, it is difficult not to bias sampling and visual impressions toward older-looking localities downvalley and younger-looking localities upvalley. Thus, because of the expectation of finding older moraines downvalley, because of the subjective nature of the judgment of the magnitude of the age-difference expected, and because of possible age differences within the Pinedale, it may be quite possible to subdivide a morainal sequence into parts inferred to be correlative with the Bull Lake and Pinedale Glaciations, even if moraines only of Pinedale age are present. Because of the fact that the age of the Pinedale end moraines may differ by more than a factor of 2, even statistically valid relative-age differences within the Pinedale could easily be misconstrued as representing the Bull Lake–Pinedale difference.

If no break judged to be of the magnitude of the Bull Lake–Pinedale difference is recognized, then yet another potential pitfall exists. Because the outer moraines are commonly identified as Bull Lake, the whole end moraine complex may be assigned to the

Bull Lake, and the Pinedale limit relegated to a canyon area or to some recessional moraines well upvalley, commonly half-way upvalley from the end moraines to the source area.

If relative-age criteria do not indicate clear age differences in an end moraine complex, considerations of regional equilibrium-line altitudes suggest that in normal cases it is more reasonable to infer Pinedale glaciers were slightly larger than Bull Lake ones, rather than that Pinedale glaciers were much smaller, say only about half the size of Bull Lake ones. Of course, special circumstances may negate this reasoning, but the similarity in size of most glaciers of Bull Lake and Pinedale age indicates that their equilibrium-line altitudes were only slightly different.

Because of its Quaternary volcanic history, the Yellowstone area offers a particularly confusing glacial history if one is strongly influenced by the Bull Lake–Pinedale paradigm. West of the park, moraines with Bull Lake characteristics have long been recognized as being well outside Pinedale ones. Therefore, it seems logical to expect Bull Lake moraines outside Pinedale ones on the northern margin of the same "Yellowstone" icemass. Consequently, in the early years of this study, I accepted the assigment by previous workers of the outer moraines in the Yellowstone valley, north of the park, to the Bull Lake Glaciation. After mapping in the Yellowstone valley, I was forced to conclude that the outer moraines have neither relative-age characteristics indicating an older age than moraines upvalley nor the general characteristics typical of Bull Lake moraines elsewhere, and that, therefore, moraines of only Pinedale age occur at the surface there. Changes in the topography of the source area due to volcanism between the Bull Lake and Pinedale Glaciations (fig. 47) furnished an explanation for the differences in relative glacier extent north and west of the park of Bull Lake and Pinedale glaciers.

In conclusion, it is obvious that Pinedale glaciers may locally have overridden and obliterated Bull Lake terminal moraines. The Pinedale-Bull Lake paradigm that I and others have used in subdividing end moraine sequences tends to ignore this hypothesis.

OUTLINE OF PRE-BULL LAKE SURFICIAL DEPOSITS

Table 1 gives the age and location of pre–Bull Lake surficial deposits in northern Yellowstone and remarks on their geologic significance. Knowledge of the pre–Bull Lake history is far more fragmentary than for the younger glaciation, but those interested in these deposits can find further information in Pierce (1973a, b; 1974a, b), Christiansen and Blank (1972), Richmond and Waldrop (1972), Brown (1961), and Boyd (1961).

BULL LAKE GLACIATION

The type area of the Bull Lake Glaciation is on the east flank of the Wind River Mountains in central Wyoming (Blackwelder, 1915; Richmond and Murphy, 1965). It is defined by well-developed end moraines outside the moraines of the last major (Pinedale) glaciation (Blackwelder, 1915, p. 324–328). Although the Bull Lake has been subdivided into an early and late stade (Richmond and Murphy, 1965; Richmond, 1965a), this report refers only to inner and outer moraines. Moreover, the Bull Lake moraines are described only at one place on the margin of the northern Yellowstone icemass. Pinedale glaciers overran and presumably obliterated Bull Lake moraines in the Yellowstone valley and in the valleys of the Gallatin River and Grayling Creek.

WEST YELLOWSTONE BASIN

A large lobe of ice of Bull Lake age advanced westward from the park into the West Yellowstone Basin, nearly filling the basin and depositing bulky end moraines (figs. 8 and 9). These moraines can be traced almost continuously over a distance of about 40 km from east of Grayling Creek to the south shore of Hebgen Lake and mostly occur between altitudes of 2,000 and 2,100 m (6,500 and 6,800 ft). This large lobe of ice appears to have flowed into the basin mostly through an area now occupied by the West Yellowstone rhyolite flow south of the present Madison Canyon (fig. 8). Glacial streaming just south of Cougar Creek and also west of Mount Haynes on the south side of the Madison River shows that ice flowed northwesterly into the basin (fig. 8). Northwesterly flow also transported obsidian erratics, probably derived from the Cougar Creek rhyolite flow, to the moraines north of Gneiss Creek.

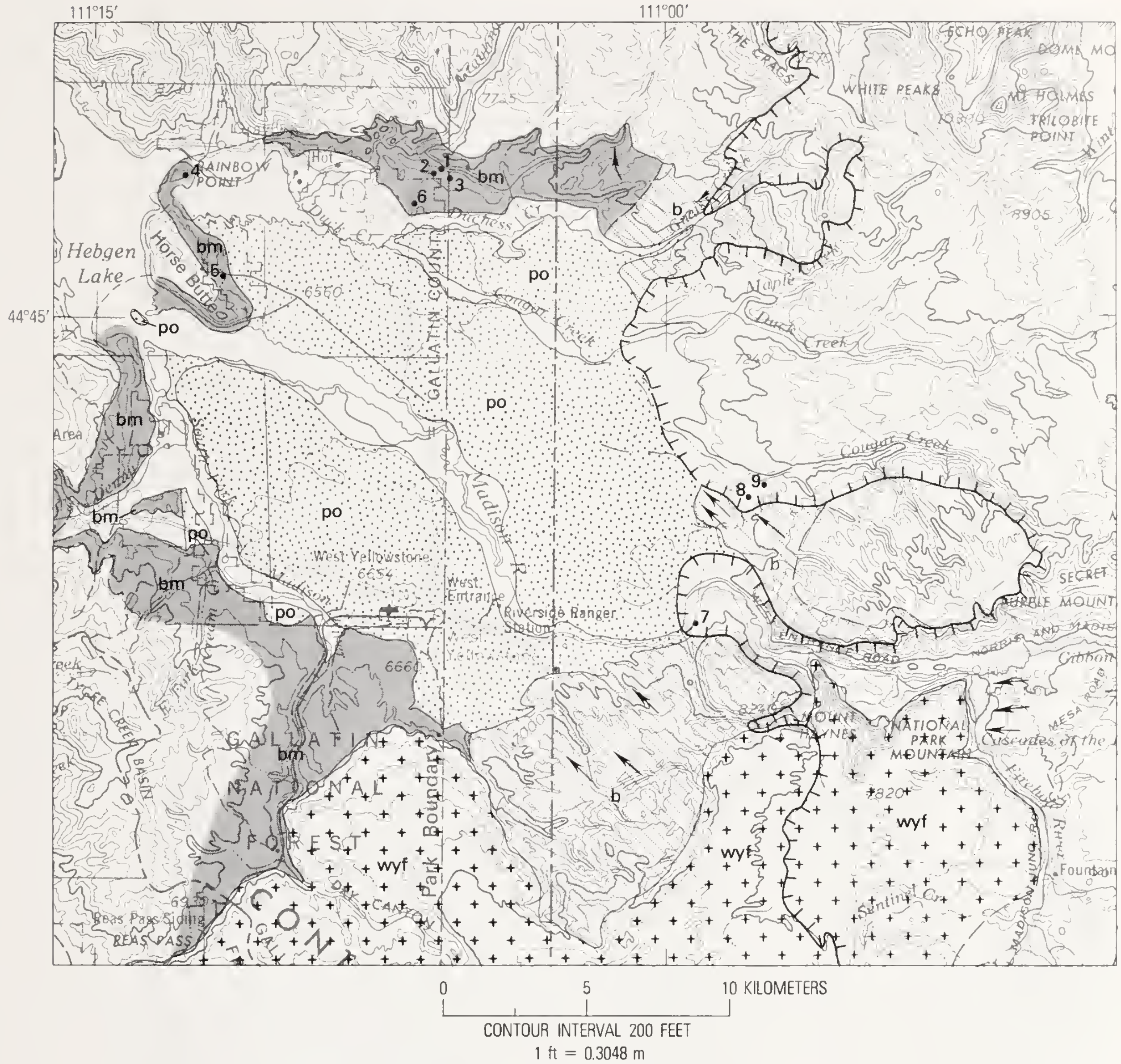
The flow of Bull Lake ice into the West Yellowstone Basin occurred prior to emplacement of the West Yellowstone rhyolite flow. South of Madison Canyon,

TABLE 1.—*Pre-Bull Lake surficial deposits in northern Yellowstone National Park*

Deposit	Age (m.y.)	Location	Remarks
Gravel beneath Junction Butte Basalt.	2	Lower third, Grand Canyon of Yellowstone River.	Indicates large stream, at present location of Yellowstone River. Position of gravel 120 m above Yellowstone River indicates that amount of net downcutting by river in last 2 m.y., but several cycles of filling and cutting are involved.
Gravel beneath Huckleberry Ridge Tuff.	2	Mt. Everts	Indicates local drainage coming from andesitic rocks of Washburn Range. Absence of reworked erratics suggests no prior extensive glaciation.
Till at Bumpus Butte and associated gravels, lake sediments, and basalts.	1.5	The Narrows near Tower Falls.	Indicates glaciers about 50 km long reached this area from Beartooth uplift. This ice first blocked and then advanced southward up the ancestral Yellowstone River, causing deposition of lake sediments, till, and associated gravel.
Pre-Bull Lake tills on (or including fragments of) Huckleberry Ridge Tuff.	<2 and >0.15	Between Gallatin and Madison Ranges.	Indicate that at least two pre-Bull Lake glaciations in this area extend further than did Bull Lake or Pinedale glaciers. The older till is deeply weathered and has no constructional morphology. The younger one retains some morainal morphology and associated outwash forms a terrace 20 m above the Gallatin River.
Diamicton beneath tuff of Notch Mountain	0.6	Saddle Mountain	If a till, glaciers may have been small because deposit occures near modern cirques. Faceted stones present, but neither striations nor erratics were noted.
Gravel beneath basalt at Deckard Flats.	0.6–0.7	Upstream from confluence of Bear Creek and Yellowstone River.	Indicates an ancient deposit of Yellowstone River of possible glacial origin. Very coarse boulder deposits at base of gravel 30 m thick suggests a glacial origin.
Gravel beneath Undine Falls Basalt.	0.6–0.7	Grand Canyon of Yellowstone River near Deep Creek.	Gravel derived from Eocene rocks of Washburn Range and local rhyolite flows indicates local drainage through area. Its position 120 m above Yellowstone River indicates that amount of net downcutting since deposition.
Glacial deposits beneath Lava Creek Tuff (not found in study area).	0.6	Northern Yellowstone area.	No definite glacial deposits were found immediately beneath the Lava Creek Tuff. This seems anomalous because the ash fall from the Lava Creek eruption (Pearlette type O) lies on Kansan tills in the western part of the midcontinent.
Gravel associated with Osprey Basalt.	0.2	Lamar River banks 30 km upstream from the Lamar Canyon.	Gravel of possible glacial origin deposited immediatly above or below basalt flows. Lamar River has cut into these deposits, but is not significantly below them.
Gravel associated with Osprey Basalt.	0.2	Sheepeater Canyon of Gardiner River.	Drainage from Gallatin Range flowed through this area. Gardiner River has cut canyon about 150 m since this time.

features produced by Bull Lake glacial scour occur immediately northwest of the West Yellowstone rhyolite flow (fig. 8; Waldrop, 1975a). Farther west Bull Lake glacial deposits extend to the rhyolite flow, but are not on it (Richmond and Hamilton, 1960; Richmond, 1964a; Waldrop, 1975a; U.S. Geol. Survey, 1972a). Pinedale ice did not advance across the northern crest of the flow, inasmuch as the top of the rhyolite there is unglaciated (fig. 8; Richmond, 1964a, fig. 115, pl. 5; Waldrop and Pierce, 1975; U.S. Geol. Survey, 1972a).

South of the Madison Canyon, Bull Lake glaciers were more extensive than Pinedale glaciers, probably owing to a combination of two related factors. First, Bull Lake but not Pinedale ice could enter the basin through an area about 15 km wide now occupied by the West Yellowstone rhyolite flow (Richmond, 1964a, fig. 115; Hamilton, 1960, p. 98; Waldrop and Pierce, 1975). Second, the center of mass of the Bull Lake icemass was probably nearer the West Yellowstone Basin than was that of the Pinedale icemass (fig. 47).



EXPLANATION

bm / b — Area covered by Bull Lake end moraines (shaded) and scour features and ground moraines (lined) beyond Pinedale limit

wyf — West Yellowstone rhyolite flow (emplaced after deposition of Bull Lake moraine)

po — Pinedale outwash (includes obsidian-sand plain)

Limit of Pinedale ice—Dashed where buried by alluvium

.5 Soil localities described in figures 10 and 11

Glacial scour features of Bull Lake age

FIGURE 8.—Map showing extent of Bull Lake glaciers near West Yellowstone, Mont. Outlines dashed where limits uncertain. Bull Lake ice extended as far west as Horse Butte, about 20 km west of limit of Pinedale ice. See figure 3 for location of this map area. Mapping of surficial deposits near West Yellowstone modified from Alden (1953), Richmond (1964a), Pierce (1973a), Waldrop and Pierce (1975), and Waldrop (1975a). Mapping of West Yellowstone flow modified from Richmond and Hamilton (1960), Hamilton (1964), and R. L. Christiansen (written commun., 1972). Base from AMS 1:250,000 Ashton, Idaho, Mont., 1955.

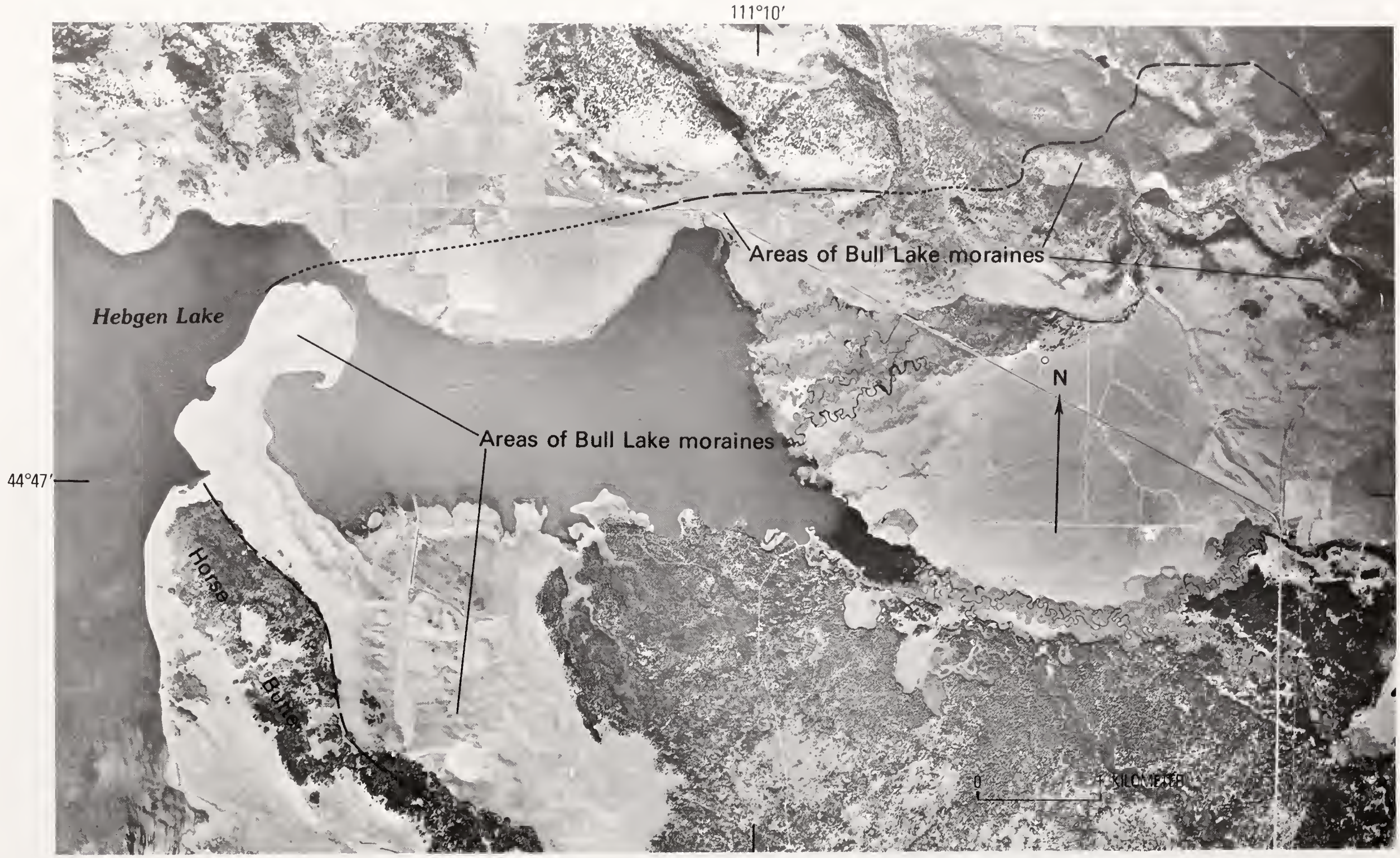


FIGURE 9.—Aerial photograph showing areas of Bull Lake moraines in the northern part of the West Yellowstone Basin. Bull Lake ice advanced westward against Horse Butte. The moraines form a well-defined morainal loop in Hebgen Lake. Their hummocky topography is still apparent in spite of a loess mantle. Poorly drained areas are still preserved in the morainal area near the bottom of the photograph. Dashed line along outer margin of moraines; dotted where inferred. Aerial photograph by AMS, 1954.

The Bull Lake ice lobe entering the West Yellowstone Basin from the southeast was joined by ice flowing westward between the south end of the Gallatin Range and the north rim of the Madison Canyon (fig. 8). Erratics from the south end of the Gallatin Range occur in Bull Lake moraines as far south as Horse Butte. These erratics were either carried directly by ice from the Gallatin Range or derived from older glacial and alluvial deposits.

Although the ice lobe south of the Madison Canyon was much larger during Bull Lake time than during Pinedale time, the amount of overflow across the divide between the south end of the Gallatin Range and the north rim of the Madison Canyon was similar during each glaciation. On The Crags (west of Mount Holmes, fig. 8), the limits of Bull Lake and Pinedale ice converge upwards and merge at an altitude of about 2,560 m (8,400 feet). South of Cougar Creek the upper limit of recognizable glaciation is thought to represent the limit of Bull Lake Glaciation. It is about 100 m (300 feet) higher than the Pinedale terminal moraines. Eastward up Cougar Creek, the Pinedale limit rises and merges with the upper limit of recognizable glaciation, suggesting that the Bull Lake was no higher than the Pinedale there. The reason for the differences in flow of Bull Lake and Pinedale ice is discussed in the concluding section of this report.

RELATIVE-AGE CRITERIA

Assignment of the moraines near West Yellowstone to the Bull Lake Glaciation is based upon the degree of contrast with nearby Pinedale moraines and their similarity to Bull Lake moraines elsewhere in the region. The main factors considered are: (1) degree of preservation of original morphology, (2) position in the moraine sequence, (3) degree of breaching by streams, (4) loess mantle, (5) weathering rinds, and (6) soil development.

The moraines of the Bull Lake Glaciation have retained most of their overall shape in plan view, as shown by the horseshoe-shaped end moraines north of Horse Butte (fig. 9). Moraine crests can be traced for kilometers. The moraines do not exhibit the pronounced hummockyness, steep slopes, or conical undrained depressions characteristic of the Pinedale moraines, but they do have the muted morphology typical of the Bull Lake (fig. 9). A few poorly drained areas between morainal highs occur just northeast of Horse Butte. The moraines still retain vestiges of their original fine-scale hummockyness (fig. 9). Slopes on the moraines are gentle, generally not exceeding 20°. The moraines are bulkier and broader than those of Pinedale age; Richmond (1965a, p. 220–221) has noted that Bull Lake moraines tend to be bulkier than Pinedale ones elsewhere in the Rocky Mountains. Because these moraines were deposited by a large, unconfined piedmont ice lobe and the Bull Lake Till is rich in fines and poor in large boulders, the original morphology of these moraines was probably never as strong as that of moraines deposited by typical Rocky Mountain valley glaciers from granitic terrain. Also, a thin loess mantle covers the till surface, burying most of the stones and smoothing the morainal morphology.

No end moraines of intermediate age occur between the Bull Lake and Pinedale moraines. In the Madison Junction quadrangle, Waldrop (Waldrop and Pierce, 1975) mapped deposits of sandy kame gravel and cemented kame gravel that overlie the West Yellowstone rhyolite flow and are mantled by surficial deposits of Pinedale age. No buried soils were found between the kame deposits and the overlying Pinedale deposits. Nonetheless, the kame deposits were assigned to the "late(?) stade of the Bull Lake(?) Glaciation" because their deposition required the presence of glaciers, and they are stratigraphically older than the mantling Pinedale deposits and younger than the West Yellowstone flow which postdates the Bull Lake end moraines. In the absence of weathering criteria indicating a Bull Lake age, I consider assignment to either the Pinedale or to a previously unrecognized glaciation equally as reasonable as assignment to the Bull Lake. The moraines assigned to the Bull Lake and Pinedale (fig. 8), however, appear to be a sequence typical of the Rocky Mountains.

The Bull Lake moraines near West Yellowstone are generally widely breached by the local streams. Commonly a segment of moraine an order of magnitude wider than the breaching stream has been removed. The width of stream-cut gaps through Pinedale moraines is only about three times greater than the width of the breaching stream.

The Bull Lake moraines are generally covered with about 0.5–1 m of windblown silt. This loess has been mixed with the underlying Bull Lake glacial deposits, and on slopes both have moved downslope by colluvial processes. The source of the eolian silt on the Bull Lake moraines may be from glacial outwash plains, arid

basins to the west, or a combination of both. Loess is widely distributed over thousands of square kilometers at the upper end of the Snake River Plain. A mantle of eolian silt is not restricted to Bull Lake deposits. Commonly 10 to 30 cm of eolian silt mantles Pinedale deposits, although this mantle is usually masked by mixing and by soil development.

In two studies, weathering rinds were measured on basalt cobbles from the Bull Lake and Pinedale moraines. In the first study, about 15 cobbles were collected from 14 sites by H. A. Waldrop from the B-horizon of the soil profile. I measured the rinds to the nearest quarter mm with a ruler. Average thicknesses for nine sites from the main morainal belt around the basin ranged between 0.7 and 1.0 mm and averaged 0.8 mm. For five sites 4 to 8 km southwest of West Yellowstone, the average thicknesses ranged from 0.95 to 1.35 mm and averaged 1.15 mm.

In a second study, S. M. Colman, aided by myself and R. M. Burke, measured rinds to the nearest 0.1 mm using a comparitor with 6× magnification. For 162 basalt stones from three localities on the morainal belt on the north side of the basin, rind thickness averaged 0.78 ± 0.19 mm (Colman, 1977). For 64 basalt stones from two localities 4–5 km southwest of West Yellowstone, the average is 0.99 ± 0.18 mm. In comparison, rinds from Pinedale end moraines average 0.40 ± 0.22 mm. These two studies show that comparable results can be obtained by the two methods; however, comparitor measurements are more precise. In addition, both studies show that the rinds are significantly thicker in the slightly higher area southwest of West Yellowstone towards Reas Pass. This increase in rind thickness probably results from increased moisture, for the vegetation there suggests moister soils, and snow survey records indicate 70 percent more moisture in the April 1 snowpack in this area than at West Yellowstone (Farnes and Shafer, 1975, p. 169–170).

On comparable parent materials, relict soils on deposits of Bull Lake age are generally better developed than on those of Pinedale age. Because of local variation in soil development it is difficult to determine which soil profiles represent the age of a given deposit. Three different studies of the soils on the Bull Lake moraines have been made. The first soils description is that of G. M. Richmond (1964a, p. 226), summarized on the left side of figure 10. A soil that I considered representative of the amount of soil development that has occurred since deposition of the moraines is shown on the right side of figure 10 and also given in Pierce (1973a, soil profile T). This profile shows somewhat better development than Richmond's generalized profile and is probably composite, with a weakly developed soil in loess mantling a better developed soil in mixed loess and till. The 7.5 YR colors in the buried B&A horizon and the clay films, and the increase in stickiness in the buried B&A horizon, in particular indicate better development than the soil in Pinedale Till or Pinedale(?) loess.

After the field studies and several drafts of this report were completed, R. M. Burke, S. M. Colman, and I examined a number of sites in the West Yellowstone Basin to better document the weathering characteristics of the Bull Lake and Pinedale moraines. I am indebted to R. M. Burke for providing the soils data shown in figure 11 (written commun., 1976). All sites except locality 5 were on flat crestal areas of moraines, or "ideal" sites.

On the right side of each graph of figure 11, the soil colors are plotted, in order of increasing intensity of oxidation, as estimated by R. R. Shroba (written commun., 1976). I have combined some colors to consolidate the diagram, combining colors recorded only once or colors where the estimate of the oxidation difference was judged to be small.

Localities 5 and 6 have the thickest loess mantle, and also have a buried textural (argillic) B horizon beneath the loess. The soil formed in the loess is less developed than the buried soil, having only a weak cambic (color) B horizon. At locality 6 the magnitude of the clay peak in the IIB2b horizon may be somewhat exaggerated, owing either to the textural variability of the parent material or to the lower permeability of the underlying compact till, which impedes deeper infiltration of clay particles. The soil profile on the Pinedale moraines is more weakly expressed than that on the surface loess, probably because the loess has more original clay and weatherable grains than the rhyolitic till.

The soil at locality 3 has the thickest profile and the greatest amount of apparent development of the sites examined. However, the location in a gentle swale, the absence of a loess mantle, and the parent material change from a diamicton to a gravel are different from the other profiles. The diamicton may be colluvium or flow till. The following explanation was given in the field by A. D. Southard (oral commun., 1971) for the solum being thicker and essentially coincident with the

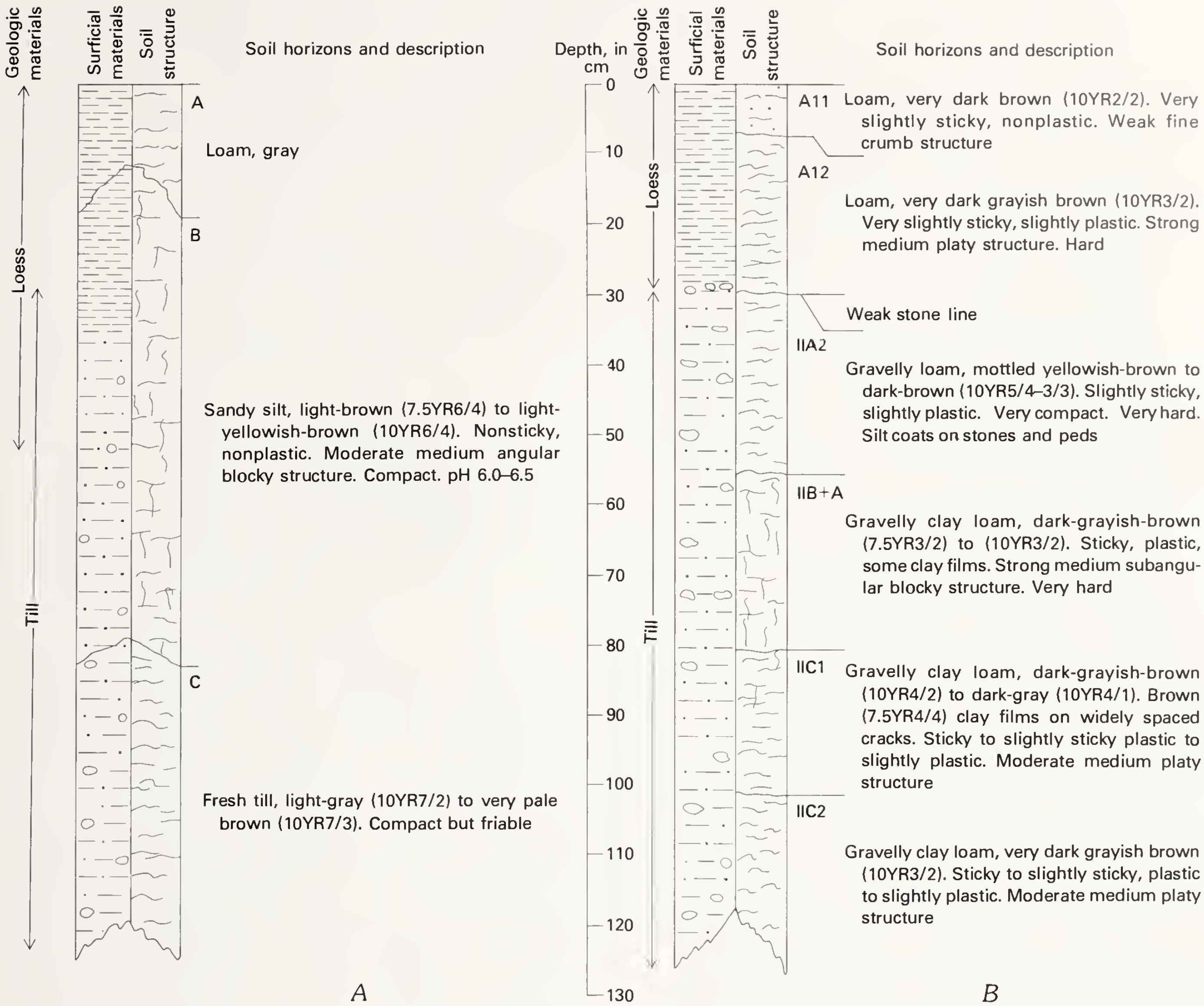
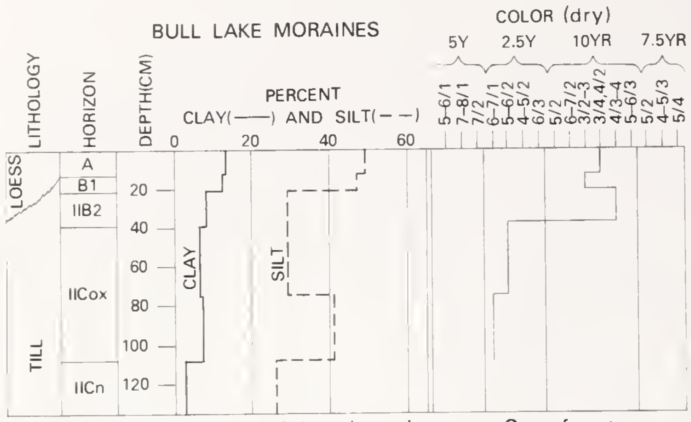
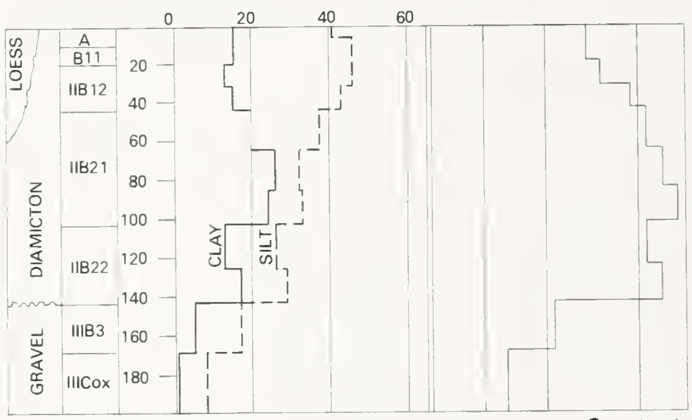


FIGURE 10.—Descriptions of soil profiles on Bull Lake moraines near West Yellowstone. Patterns in surficial-materials column indicate stones (circles), sand (dots), and silt and clay (dashes). *A*, Relict soil as generalized from Richmond (1964a). *B*, Soil on moraine crest at locality no. 2 (fig. 8). Vegetation, grassland with widely scattered trees. Colors are for moist soil. A buried soil occurs below 30 cm. The modern soil is developed through a mantling loess into the buried soil. Measured with Clifford Montagne in 1970 and A. D. Southard in 1971.

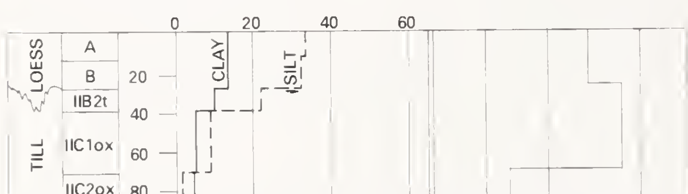
diamicton layer. With the annual snowmelt, the entire diamicton layer is wetted to capacity before water can move downward into the gravel. In thicker deposits of similar texture, capillary tension would have facilitated movement of water to greater depths and the moisture content of the upper 1.4 m would be less. Thus, the diamicton layer is wet and dried, or "worked" as a unit, thereby giving a unity to its oxidation state and clay content. This seasonal wetting and the well-drained gravel beneath the diamicton facilitated stronger development of oxidation colors. The clay may either be original, be from infiltrated loess, or result from colluviation of loess-rich soil between two moraine crests.



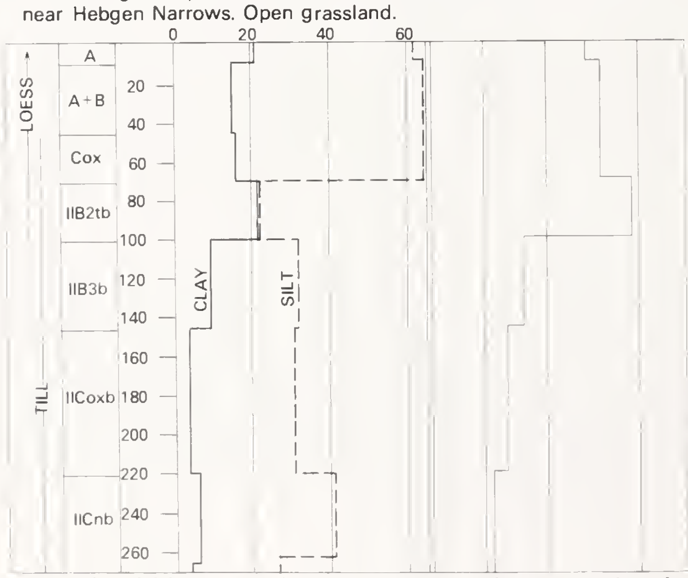
Loc. 1. Roadcut through broad moraine crest. Open forest.



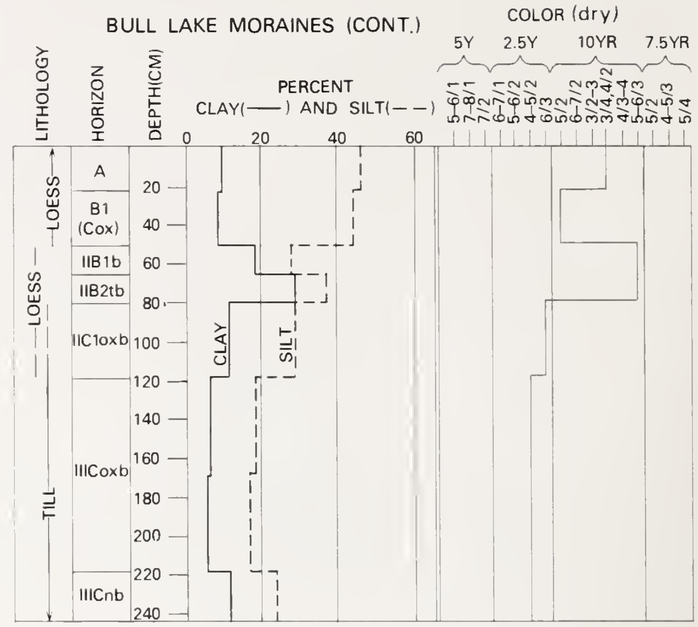
Loc. 3. Backhoe pit in colluviated swale along moraine crest. Grassland with widely scattered trees.



Loc. 4. Pit dug in very broad moraine crest of horseshoe-shaped terminus near Hebgen Narrows. Open grassland.

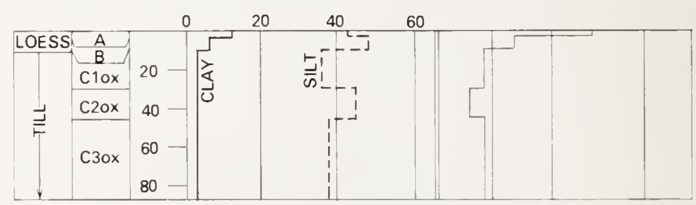


Loc. 5. Subdivision roadcut in gently sloping moraine on east side of Horse Butte. Soil affected by colluviation in upper 1.5 m. Grassland.

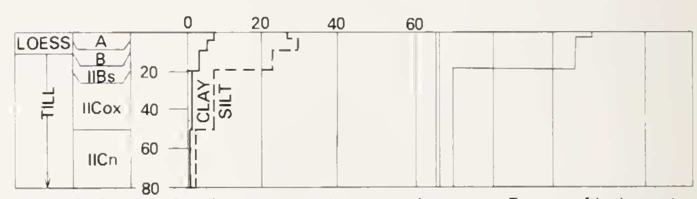


Loc. 6. Broad moraine crest adjacent to obsidian-sand plain. Grassland.

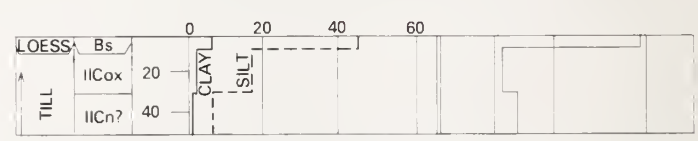
PINEDALE MORAINES



Loc. 7. Madison terminus. Wide crestal area on small moraine. Forest of lodgepole pine.



Loc. 8. Cougar Creek area. Bouldery moraine crest. Forest of lodgepole pine.



Loc. 9. Cougar Creek area. Bouldery moraine crest inside that of loc. 8. Forest of lodgepole pine.

The soils on the Pinedale moraines at the mouth of the Madison Canyon and just to the north along Cougar Creek are developed in such different parent material that comparison with the Bull Lake soils is misleading. Where post-Pinedale soils are developed in parent material more comparable to the Bull Lake Till, such as on both sides of the Gallatin Range (discussed in the section on the Gallatin Valley glacier), a cambic B horizon (or Cox) extends to a depth of about 0.8 m; no evidence was found in the field of a textural (or argillic) B horizon. These areas were deglaciated as recently as about 15,000 years ago. (See "Age of the Deckard Flats readjustment.")

DISCUSSION OF AGE ASSIGNMENT

The moraines flanking the West Yellowstone Basin were assigned to the Bull Lake Glaciation or to ages considered correlative with the Bull Lake by all studies completed before obsidian-hydration dating of these moraines. Alden (1953, p. 176) mapped these moraines and assigned them to the "Iowan or the Illinoian glaciation," a designation he considered (p. 86) equivalent to "Blackwelder's (1915, p. 324-328) Bull Lake stage of glaciation." Later studies by Richmond (1964a, 1965a, 1965b) subdivided these moraines into the early and late stades of the Bull Lake Glaciation.

FIGURE 11(facing page).—Grain sizes and colors of soil profiles on Bull Lake and Pinedale deposits in the West Yellowstone Basin. These profiles were described and analyzed by R. C. Burke (written commun., 1976) with assistance in the field by myself and S. M. Colman. Numbered localities appear in figure 8. The Bull Lake moraines (localities 1, 3-6) are mantled by loess of Pinedale age and perhaps some older loess, but only at localities 5 and 6 is it clear that the upper loess buries a soil with a textural B horizon. Even though localities 1 and 4 are on broad moraine crests, the soils are less well developed, possibly because they have been truncated. See text for discussion of locality 3.

The post-Pinedale profiles (localities 7-9) cannot be compared directly with the post-Bull Lake profiles because the till of Pinedale age is rather inert rhyolite sand and silt, whereas the till of Bull Lake age has more primary clay and weatherable rock and mineral grains, and primary clay is also available from the loess mantle on the Bull Lake deposits. Colors arranged in order of increasing oxidation. Clay and silt, percent of $<$2mm fraction.

Witkind (1969, p. 55) assigned these moraines to the Bull Lake, as did Waldrop (1970, 1975a), Waldrop and Pierce (1975), Pierce (1973a), and Fred Nails (oral commun., 1972). On the surficial Geologic Map of Yellowstone National Park (U.S. Geological Survey, 1972a) compiled by G. M. Richmond, K. L. Pierce, and H. A. Waldrop, these moraines were assigned by G. M. Richmond to the *late* stade of the Bull Lake Glaciation.

Since about 1960, the Bull Lake has been widely accepted as being of early Wisconsin age (Richmond, 1965a, table 2; 1970, fig. 6; Scott, 1965, table 1; Morrison, 1965, table 1). In fact, this correlation became so well accepted that when the Bull Lake-Pinedale sequence was correlated with the Riss-Würm sequence of the Alps, the Riss was therefore inferred to be of early Wisconsin age (Richmond, 1970a). This implies that no interglacial time or soil formation of interglacial magnitude,[1] such as the Sangamon, occurred between either the Riss and Würm Glaciations, or the Bull Lake and Pinedale Glaciations.

The first numerical dating of moraines assigned to the Bull Lake Glaciation was orally presented at the 1972 meetings of the American Association for Quaternary Research, and published 4 years later (Pierce and others, 1976). The Bull Lake moraines near West Yellowstone are stratigraphically older than the West Yellowstone flow now dated as 117,000±8,000 years old (weighted means of 4 dates, $2\sigma_x$, J. D. Obradovich, written commun., 1979, using K-Ar constants of Steiger and Jäger, 1977). Obsidian hydration measurements on glacially abraded stones from the moraines compared with cooling cracks on the K-Ar

[1]The concept of the duration of interglacial times has undergone major revision recently, most convincingly as a result of studies of the marine Pleistocene record. Rather than interglacial climates being "normal," punctuated by shorter glacial episodes, we now know that times as warm as the present have averaged only about 10,000 years in duration, and that 90 percent of the last half-million years has been more glacial than the present (Emiliani, 1972). Thus, in the stratigraphic record, soils and other evidence of the last interglacial need not indicate an order of magnitude more time than the post-glacial or Holocene. First the peculiar concept of soil-forming intervals needs to be placed in proper perspective (Birkeland and Shroba, 1974); then the soil formed during the *culmination* of the last interglacial should look rather similar to the post-Pinedale soil, because this interval (isotope stage 5e) was similar to the Holocene, in both duration and extent of global deglaciation (Emiliani, 1972; Shackleton and Opdyke, 1972, p. 46). Of course, the soil formed during all of isotope stage 5, which may represent the full time-span of the last interglaciation, would probably be significantly better developed than the post-Pinedale soil, for this interval lasted about 50,000 years, from about 125,000 to about 75,000 years ago.

dated West Yellowstone and Obsidian Cliff rhyolite flows (1977 constants) indicate that the average age of glacial cracking is about 150,000 years old, with most cracks dating between 135,000 and 160,000 years old. These ages are rounded off from a graphical solution like that in Pierce and others (1976, fig. 4) but modified by increasing the K-Ar control ages by 2.6 percent.

This age is older than the last interglacial of the marine record, and it indicates that the moraines are of pre-Wisconsin age. Following this first age-dating of moraines assigned to the Bull Lake Glaciation, their Bull Lake age was questioned by Richmond (1976, p. 373), who stated that "the surficial weathering characteristics of till 5 [same map unit as Bull Lake moraines of this report] suggest correlation with the broad smooth moraines of the late stade of the Sacagawea Ridge Glaciation, but do not exclude the possible correlation with the broad smooth moraines of the early stade of the Bull Lake as locally developed at Bull Lake. This distinction is easily made in the Wind River Range because moraines of the Sacagawea Ridge Glaciation are related to a much higher outwash terrace than those related to the Bull Lake moraines."

The properties of the moraines flanking the West Yellowstone Basin fit rather well the relative-age and stratigraphic criteria generally recognized as the basis for the assignment of Bull Lake age (Blackwelder, 1915; Richmond, 1962a, 1965a, 1965b; Richmond and Murphy, 1965). The basic stratigraphic framework for the younger part of the Quaternary sequence in the Rocky Mountains has been outlined as follows (Richmond, 1962a, 1965a, 1970):

Immature zonal, or moderately developed soil
Deposits of Pinedale age
Mature zonal, or strongly developed soil
Deposits of Bull Lake age, late stade
Mature zonal, or strongly developed soil
Deposits of Bull Lake age, early stade
Very strongly developed soil, may be partly stripped.

The deposit-and-soil sequence on the Bull Lake moraines near West Yellowstone is shown in profiles 5 and 6 (fig. 11), and is as follows: (1) a moderately developed soil in Pinedale loess, (2) less weathered Pinedale loess, (3) strongly developed soil in till, and (4) unweathered till. Using the above stratigraphic framework, the till would fit the position of the late stade of the Bull Lake Glaciation.

An important parameter reflecting the degree of soil development is the accumulation of clays in the B horizon. A simple way to quantify the amount of clay increase is shown in table 2, where the increase in clay is calculated as an equivalent thickness of clay in centimeters. The moraines flanking the West Yellowstone Basin have a clay increase estimated in the range of 1.1 to 3.3 cm (averaging 2.2 cm), and the Bull Lake deposits from the other areas listed have a clay increase estimated in the range of 1.1 to 2.7 cm (averaging 2.1 cm). These data suggest the Bull Lake assignment of the moraines near West Yellowstone is reasonable.

In terms of end-moraine sequence, the only moraines still preserved that are possible candidates for the Bull Lake in the West Yellowstone Basin are those flanking the basin. In the Rocky Mountains, Bull Lake moraines are normally found beyond Pinedale moraines. Pinedale glaciers could have overridden and obliterated Bull Lake moraines, but if so, this requires an explanation separate from the one advanced near the end of this report ("Comparison of Bull Lake and Pinedale Glaciations") for the absence of Bull Lake moraines in the Yellowstone valley and the presence of moraines of Bull Lake character at West Yellowstone.

Weathering rinds can be used to compare the West Yellowstone sequence with other typical Rocky Mountain sequences. Weathering rinds on basalt cobbles from the Bull Lake moraines of the West Yellowstone Basin are twice as thick as those on nearby Pinedale moraines. (See section titled "Relative-age Criteria.") At McCall, Idaho (Schmidt and Mackin, 1970; Fryxell and others, 1965, p. 91), weathering rinds on basalts from Bull Lake moraines are four times thicker than those from the Pinedale moraines (Colman, 1977, fig. 11). Thus, the weathering rind data suggest that the West Yellowstone moraines are no older than moraines of typical Bull Lake character at McCall.

Richmond (1976) appeared to favor correlation of the *moraines* flanking the West Yellowstone Basin with those of the Sacagawea Ridge Glaciation. Type deposits of the Sacagawea Ridge Glaciation are present only in stratigraphic section on the west side of Dinwoody Lake, Wyo., where they are buried beneath younger glacial deposits (Richmond, 1962b, 1964b). According to Richmond (oral commun., Feb. 22, 1977), in the type area, *moraines* of Sacagawea Ridge age are 2.4 km east of the north end of Dinwoody Lake, approximately centered on sections 22, 23, 26, and 27, T. 5 N., R. 5 E., on the Wilderness, Wyo., 7½-minute quadrangle. These moraines are graded to the "upper" Sacagawea Ridge terrace, 150 m (500 ft and bounded by the 6,800-ft contour) above the Wind River (Richmond, 1964b; 1976, p. 358; oral commun., 1977). The "upper" but not the "lower" Sacagawea Ridge terrace is well represented in the area of the type Sacagawea Ridge Till. Six kilometers upvalley from the "type" Sacagawea Ridge moraines is a deposit of mainstream terrace gravel from the Wind River containing volcanic ash (Richmond, 1976, p. 360)

TABLE 2.—*Estimated clay increase in B-soil horizon in deposits of Pinedale, Bull Lake and pre-Bull Lake age*

Place	Age and locality (fig. 8)	Soil Horizon	Thickness in cm	Percent clay	Percent original clay estimated [1]	Percent clay increase	Percent of sample <2 mm	Percent of clay increase, whole sample	Equivalent thickness of clay [2] in cm	References and remarks
West Yellowstone, Montana.	Pinedale, loc. 8.	B	6	6	4–6	0–2	70	0–1.4	(0–0.1)	R. C. Burke, written commun., 1966.
		IIBs	10	4	2–4	0–2	60	0–1.2	(0–0.1)	
									(0–0.2)	
	Bull Lake, loc. 5.	IIB2t	30	22	12–17	5–10	70	3.5–7.0	(1.0–2.1)	Do.
		IIB3	45	9	5–8	1–4	65	0.65–2.6	(0.3–1.2)	
									1.3–3.3	
	Bull Lake, loc. 6.	IIB16	15	19	15–17	2–4	~80	1.6–3.2	(0.2–0.5)	Do.
		IIB2tb	15	29	15–20	9–14	~70	6.3–9.8	(0.9–2.8)	
									1.1–3.3	
Rocky Mountains; Wyo., Colo., N.M.	Bull Lake, (average of four areas).	Bt	40	17	5–9	8–12	~50	4–6	1.6–2.4	R. R. Shroba (1976), includes type Bull Lake deposits.
	Pre-Bull Lake (1 locality).	Bt	90	28	5–7	21–23	~60	12.6–13.8	11–12	Do.
Denver, Colo., area, type Louviers Alluvium.	Bull Lake, loc. 8.	B2t	30	18	8–10	8–10	~90	7.2–9.0	2.2–2.7	Machette and others (1976a, b). Both localities on type Louviers Alluvium.
	Bull Lake, loc. 9.	B22t	20	26	15–19	7–11	~80	5.6–8.8	(1.1–1.8)	
		IIB3t	18	12	6–8	4–6	~50	2–3	(0.4–0.5)	
									1.5–2.3	

[1] This estimate is based on profile descriptions and plots of particle sizes with depth. Because the estimate requires subjective judgments, it is the least reliable part of the calculation.
[2] Numbers in parentheses indicate subtotals where B horizon is subdivided.

discovered by J. F. Murphy (his locality number C-14-53). The ash was subsequently recollected by R. E. Wilcox (loc. 65W74). Richmond (1976, p. 360) reported that "According to G. Izett (oral commun., 1975) this ash has the petrographic characteristics of the Pearlette type-O volcanic ash." Superhydration dating studies by V. Steen McIntyre (Richmond, 1976, p. 360) also indicate an age similar to the type-O from Wascana Creek, Saskatchewan. The Pearlette type-O is about 600,000 years old, and related to the eruption of the Lava Creek Tuff from the Yellowstone area (Naeser and others, 1973). The ash locality discussed above occurs about 140 m above the Wind River, slightly less than the main Sacagawea Ridge terrace in the Dinwoody Lake area. Assuming the ash is correctly identified as related to the Lava Creek eruption, the Wind River was *at or below* the level of the Sacagawea Ridge terrace 600,000 years ago. The type Sacagawea Ridge moraines, which are graded to this terrace, are probably about 600,000 years old.

The moraines flanking the West Yellowstone Basin are about 150,000 years old. Consequently correlation of these moraines with the Sacagawea Ridge Glaciation seems unreasonable, for "type" Sacagawea Ridge moraines are more than 400 percent older than those near West Yellowstone. On the other hand, Richmond maintained (1977b) that the oldest part of the Bull Lake Glaciation is about 115,000 years old, only 25 percent younger than the obsidian-hydration age of the West Yellowstone moraines. The relative-age criteria available for delimiting Bull Lake deposits are not, in my opinion, capable of distinguishing a 25 percent difference in age, but they should be capable of distinguishing a 400 percent difference.

The foregoing evidence seems to strongly favor retention of the Bull Lake age assignment for the moraines at West Yellowstone, and to strongly argue against introduction of a Sacagawea Ridge age assignment. I do not wish to imply that all moraines assigned to the Bull Lake Glaciation are about 150,000 years old; indeed some may be significantly younger and others significantly older. Because no age differences are apparent within the main belt of Bull Lake moraines at West Yellowstone and because of the ambiguities of previous usage of "early" and "late" designations (Pierce and others, 1976, p. 709), assignment to one subdivision or the other does not seem appropriate at this time.

GLACIERS ON WESTERN MOUNT COWEN

No Bull Lake end moraines are recognized for the northern Yellowstone outlet glacier, presumably because they were overrun by the Pinedale outlet glacier. Nevertheless, it is possible to compare Bull Lake and Pinedale moraines in the Yellowstone valley, because just downvalley from the terminus of the Pinedale outlet glacier, Bull Lake and Pinedale moraines deposited by local valley glaciers are preserved. Because all the glaciers between Elbow Creek

and Pine Creek headed on Mount Cowen, they are referred to collectively as the western Mount Cowen glaciers (fig. 12).

End moraines of the western Mount Cowen glaciers as mapped by Horberg (1940) and Montagne (1968, 1970, written commun., 1973) indicate that the Bull Lake and Pinedale glaciers were nearly the same size. Along Strawberry Creek (fig. 12), the Pinedale glacier was larger, but Bull Lake moraines were not obliterated because the Bull Lake glacier flowed in a northwesterly direction as it entered the Yellowstone valley, whereas the Pinedale glacier flowed northeasterly. Here the Bull Lake moraines can be subdivided into an outer and inner moraine; the outer moraine bifurcates to form two ridges.

The Bull Lake moraines of the western Mount Cowen glaciers are more subdued, less bouldery and hummocky, and broader crested than those of Pinedale age. Exposures along a logging road across the Bull Lake moraines of Strawberry Creek reveal that oxidation extends to a depth of more than 1.2 m, and a slightly clayey B-horizon is present. The surface boulders of granitic rocks on Bull Lake moraines are generally deeply pitted and lack glacial polish.

The relative difference in the amount of fault offset along range-front faults that bound each side of the

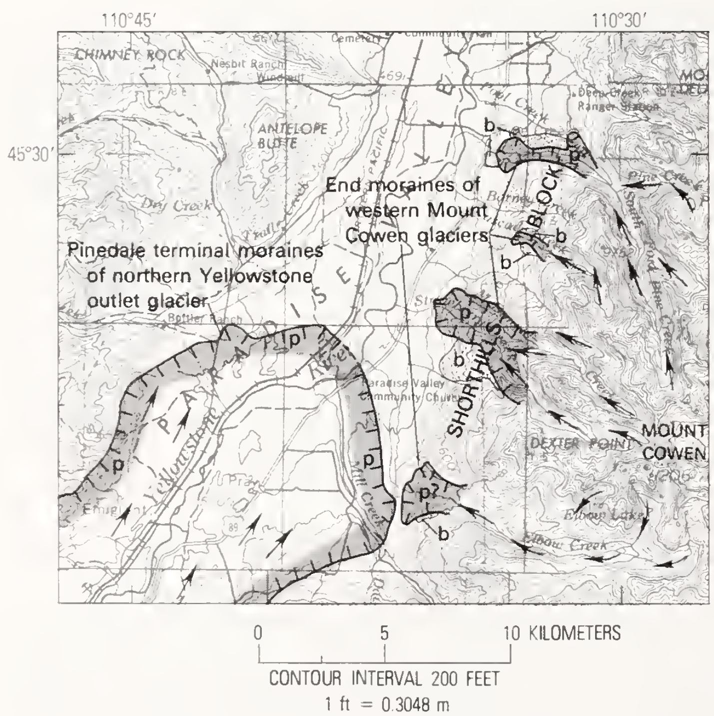


FIGURE 12.—Map of Bull Lake and Pinedale end moraines of western Mount Cowen glaciers in Yellowstone valley (modified from Horberg, 1940; Montagne, 1968, 1970, written commun., 1973). Two distinct ages of moraines are recognizable for these local glaciers, b, Bull Lake, and p, Pinedale, but a similar contrast is not apparent among the moraines of the northern Yellowstone outlet glacier. Base from AMS 1:250,000 Bozeman, 1965.

Shorthills Block (fig. 12) of Montagne (1968) is an indication of the relative age of Bull Lake and Pinedale moraines. Field studies and preliminary photointerpretation by John Montagne, W. B. Hall, and myself suggest that along the fault system offset of Bull Lake moraines is roughly five times greater than offset of Pinedale moraines. If the rate of fault movement is roughly constant for the time intervals involved, this relative offset provides an index of the relative ages of these deposits.

BULL LAKE DEPOSITS WITHIN THE AREA COVERED BY PINEDALE ICE

At a few localities in the northern Yellowstone area, glacial deposits are tentatively correlated with the Bull Lake Glaciation, largely because they predate Pinedale deposits. The age of these deposits is inferred primarily from their relation to the deposits and flow patterns of Pinedale ice. My assignment of these deposits to the Bull Lake first required an understanding of the Pinedale Glaciation.

NORTHERN YELLOWSTONE OUTLET GLACIER

Evidence is meager of a glaciation more extensive than the Pinedale terminus at the Eightmile moraines (pl. 1*A*). In many months of fieldwork John Montagne (oral commun., 1971) found no evidence of any glaciation more extensive than the Pinedale on the northwest side of the valley. However on the southeast side a few deposits containing erratics (not necessarily of glacial origin) may indicate that a larger outlet glacier existed during pre–Bull Lake time (Montagne, 1968, 1970).

The Bull Lake and Pinedale moraines of the western Mount Cowen glaciers impose two constraints on the size of the northern Yellowstone outlet glacier during Bull Lake time. First, no Bull Lake outlet glacier could have advanced downvalley as far as Strawberry Creek, for it would have obliterated the local Bull Lake moraines there (fig. 12). Second, the degree to which the local Bull Lake moraines retain constructional topography makes it quite improbable that moraines of the same age deposited by an outlet glacier would not be recognized.

A sizeable northern Yellowstone outlet glacier surely existed in Bull Lake time, but the Pinedale was simply more extensive. Buried deposits of possible Bull Lake age found in a few places seem to corroborate this explanation. Thick gravels beneath the large Pinedale outwash fan (fig. 16, pl. 1*A*) commonly show a distinct morphologic break about halfway between the river level and the surface of the fan. The gravels below the break may be Bull Lake in age.

Deposits of inferred Bull Lake age are exposed in bluffs along the Yellowstone River 3 km upstream from Emigrant, Mont. The section is:

	Thickness (meters)
Pinedale till, not exposed at top of section, but forms the pronounced morainal topography farther upslope.	
Gravel, tan; probably fan gravel from Emigrant Creek; more angular, finer grained, and not as well sorted as that below	6
Gravel, gray, well-sorted; Yellowstone River outwash of inferred Bull Lake age	4.5
Covered, probably gravel	10
Lake sediments, gray, laminated, compact silt with some sand	6

The well-sorted gray gravel looks like normal outwash, and by stratigraphic position it is either older Pinedale or late Bull Lake in age. In either case the underlying lake sediments probably date from the Bull Lake Glaciation. They probably formed when a Bull Lake morainal dam located several km downstream backed water up into this area. Pinedale ice subsequently overrode and compacted these lake sediments.

NORTHERN YELLOWSTONE NATIONAL PARK

As explained in figure 13, the contrast in Gardners Hole between the erratics in the surface till and those in a buried Bull Lake(?) Till suggest that the Bull Lake glacier was smaller than the Pinedale glacier there.

In the upper Lamar drainage, lake sediments and associated gravel and till that occur beneath deposits of Pinedale age are probably of Bull Lake age. Their position and character indicate an origin similar to that of nearby Pinedale sediments, which accumulated when base level was raised by ice-damming downstream.

TRANSPORT OF BEARTOOTH ERRATICS SOUTHWARD PAST THE WASHBURN RANGE

At some time after deposition of the Bull Lake moraines near West Yellowstone and before full-glacial Pinedale time, ice from the Beartooth uplift pushed southward between Mount Washburn and Specimen Ridge to the Canyon Village area carrying erratics of Precambrian rocks (fig. 14). Although not abundant, these erratics are conspicuous on the rhyolite terrain. Since first recognized by Holmes (1881), these erratics have served to prove extensive glaciation of the northern Yellowstone area. Although these erratics do require large glaciers in northern Yellowstone, such glaciers were probably not as large as those during Pinedale full-glacial time (pl. 1*A*). Under Pinedale full-glacial conditions, flow was northward across the Washburn Range (pl. 1*A*), but when icecaps on the

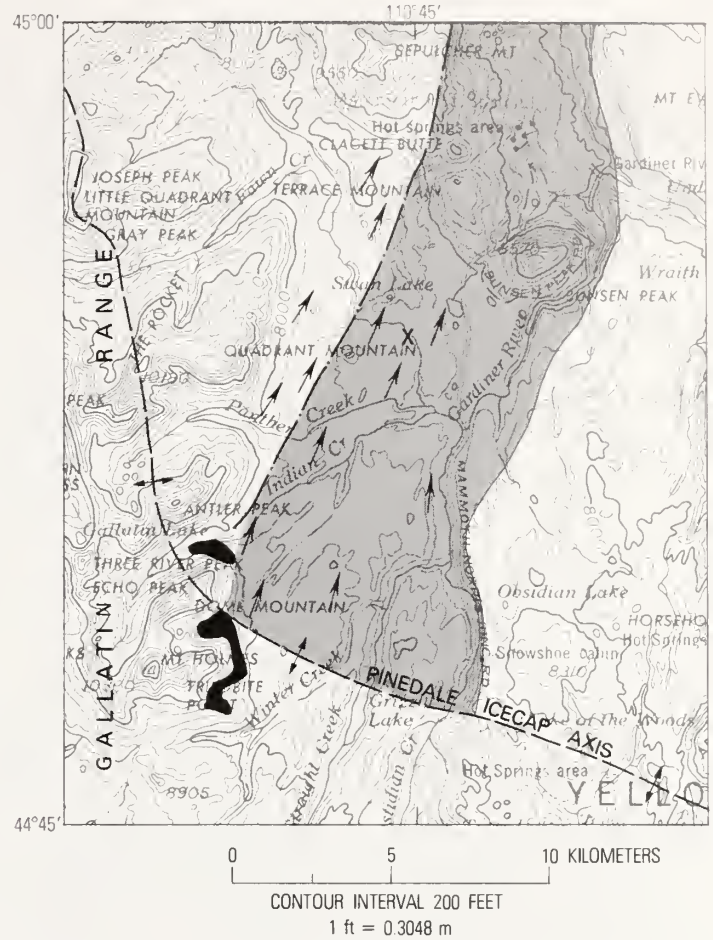
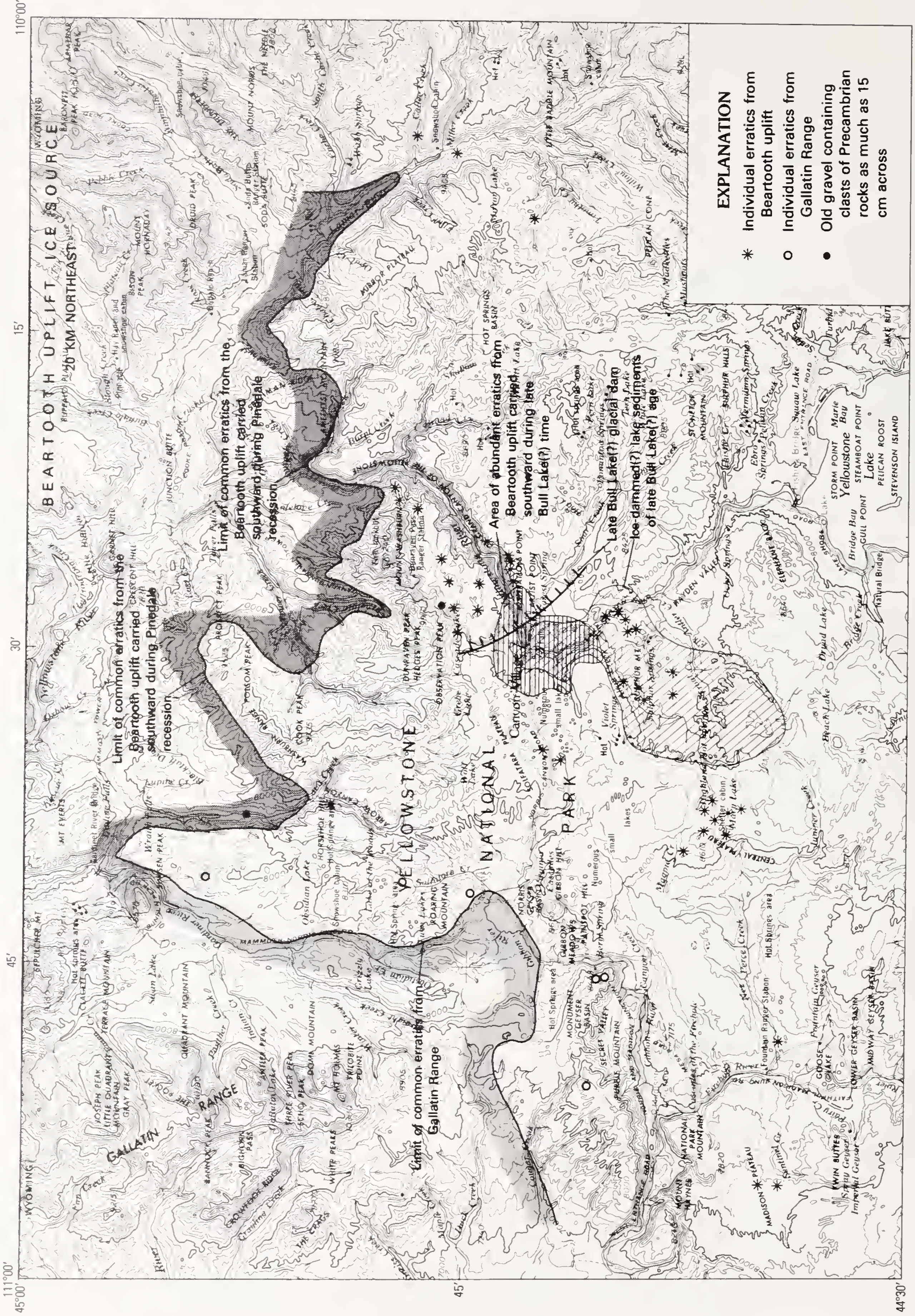


FIGURE 13.—Features east of the Gallatin Range suggesting Bull Lake glaciers there to be smaller than those of Pinedale age. Arrows represent flow direction of Pinedale ice as indicated by drumloid topography (frontispiece). Shaded area, Precambrian erratics in Pinedale Till; dash dot line, northwest limit of Precambrian erratics in Pinedale Till. Black area, sources of Precambrian erratics. Erratics indicate that Gallatin Range ice from the Indian Creek valley was pushed northward by an icecap to the east (frontispiece). X indicates where Pinedale Till overlies a smeared-out soil(?) developed on till inferred to be of Bull Lake age. At the X and to the dash-dot line 1.5 km to the northwest, Pinedale Till contains erratics of Precambrian rocks, whereas the underlying Bull Lake(?) Till does not. Instead, it contains debris from a more westerly and less southerly source than the Pinedale. This difference in erratics suggests that the Pinedale icecap east of the Gallatin Range was larger than the Bull Lake one. Base from AMS 1:250,000 Ashton, Idaho, Mont., 1955.

Yellowstone plateau were small or absent, both before and after Pinedale full-glacial time, glacier flow was southward from the Beartooth uplift.

Previous explanations of these erratics mostly assumed a simple history of southward transport during the last major glaciation (Holmes, 1881; Howard, 1937, pl. 15), with the noteworthy exception of Alden (1928, p. 63), who considered the main glacial flow in the area to have been northward and postulated ice-rafting to explain the erratics.


110°00'
45°00'
111°00'
15'
30'
45'
44°30'
45'
BEARTOOTH UPLIFT ICE SOURCE
20 KM NORTHEAST
GALLATIN RANGE
YELLOWSTONE NATIONAL PARK
Limit of common erratics from the Beartooth uplift carried southward during Pinedale recession
Limit of common erratics from the Beartooth uplift carried southward during Pinedale recession
Limit of common erratics from Gallatin Range
Area of abundant erratics from Beartooth uplift carried southward during late Bull Lake(?) time
Late Bull Lake(?) glacial dam
Ice-dammed(?) lake sediments of late Bull Lake(?) age
EXPLANATION
Individual erratics from Beartooth uplift
Individual erratics from Gallatin Range
Old gravel containing clasts of Precambrian rocks as much as 15 cm across
0 10 20 KILOMETERS
CONTOUR INTERVAL 200 FEET
1 ft = 0.3048 m

It should be noted that a conglomerate exposed over a limited area south of the Washburn Range provides a source for small erratics of Precambrian crystalline rocks and quartzite. One sizeable exposure 5 km northeast of Canyon Village (fig. 14; Howard, 1937, table 10; Pierce, 1974a; Prostka, Ruppel, and Christiansen, 1975a) contains no clasts larger than 20 cm in diameter. A number of the small erratics of Precambrian crystalline rocks probably came from this source, but boulders more than 30 cm across probably were carried southward by ice from the Beartooth uplift.

More than 100 erratics of Precambrian rocks have been found in the Canyon Village and Hayden Valley areas, and on the slopes of Mount Washburn below an altitude of about 2,600 m (8,550 ft) (fig. 14; Howard, 1937, fig. 15, table 10, p. 85–90; Pierce, 1974a; Richmond and Waldrop, 1975, Richmond, 1977a). The erratics are much less common than in Pinedale recessional moraines from the Beartooth uplift. Oddly, few erratics can be found between the Canyon Village area and their source (fig. 14). This distribution of erratics suggests a history more complicated than a simple southward advance during the most recent glaciation.

The glacier that dammed the Yellowstone River forming Hayden Lake of Howard (1937, p. 98, pl. 15; fig. 14) probably also transported most of the erratics south of the Washburn Range. As shown by Howard (1937, pl. 15), a glacier advancing from the north formed a northwest-southeast-trending dam just southwest of Canyon Village; the sediments deposited in the lake that resulted are assigned a "late Bull Lake" age (U.S. Geological Survey, 1972a; Richmond, 1977a). These lake sediments are considerably younger than the Bull Lake moraines near West Yellowstone dated as about 150,000 years old, for they overlie the Hayden Valley rhyolite flow (Richmond, 1977a) dated as about 100,000 years old (J. D. Obradovich, written commun., 1974). Howard (1937, p. 99) suggested that the "scattered granite boulders in these clay-silts were probably ice-rafted to their present positions" from the glacier that dammed the lake.

Granitic erratics are concentrated just southwest of the inferred ice dam in the area of lake sediments of "late Bull Lake" age (fig. 14). For example, 30 erratics were found in a 90-minute search in the area of lake sediments about 3 km south of Canyon Village where Howard (1937, pl. 15) showed his ice dam for Hayden lake, and 5 were found in 45 minutes in the area of lake sediments northwest of the Lower Falls. In addition, erratics of andesitic volcanic rocks, probably from the Washburn Range (also to the north), are common both on the surface of these lake sediments and as stone bands in roadcuts. In contrast, no Precambrian erratics were found during 5 days of traverses over 200 km² east of the Yellowstone River between the area of lake sediments and Deep Creek (fig. 14). This is a remarkable lack, because the latter area is closer to the source of erratics and the ice that transported them southward must have passed across that terrain. Pinedale glaciers apparently removed the erratics in this area, and probably reworked most of the erratics found south of the Washburn Range. Although Pinedale ice flowed northward over the Washburn Range, apparently only minor scour occurred at the ground surface south of the range (fig. 27). However, Pinedale ice has scoured to bedrock and apparently removed nearly all the erratics of Precambrian rocks from the area between the Washburn Range and Specimen Ridge, because no topographic obstacle impeded northward flow.

Southwest of Canyon Village, erratics of Precambrian rocks are found in the Mary Mountain area (Central Plateau) at altitudes as high as 2,600 m (8,550 ft) (Howard, 1937, table 10). West of Mary Mountain, one large erratic occurs along Nez Perce Creek just east of the Lower Geyser Basin, and two large erratics occur near the western limit of Pinedale ice west of the geyser basin (Waldrop and Pierce, 1975). If the erratics were carried southward from the Beartooth uplift by direct flow of Pinedale ice, normal relations between ice thickness and surface slope (fig. 48; pl. 4) would require an ice surface well above 3,400 m (11,000 ft) over Specimen Ridge and above 3,200 m (10,500 ft) on Mount Washburn. These altitudes for southward-flowing ice seem excessive, for erratics of Precambrian rocks are absent above 2,750 m (9,000 ft) on Specimen Ridge and above 2,620 m (8,600 ft) in the Washburn Range, and striations above these altitudes indicate flow to the northwest, rather than southwest (pl. 1*A*). Thus, these erratics were probably carried south of the Washburn Range by late Bull Lake or early Pinedale ice and then westward by Pinedale ice.

The transport of these erratics to south of the

FIGURE 14.—Transport of Beartooth erratics southward past the Washburn Range. Before Pinedale full-glacial time, Beartooth ice advanced southward and blocked the Yellowstone River near Canyon Village, forming "Hayden Lake" in the Hayden Valley area (Howard, 1937, pl. 15). The area of abundant erratics from the Beartooth uplift indicates this glacier terminated in a lake just east of Canyon Village. The resulting lake sediments are assigned to the "late stade" of the Bull Lake Glaciation (U.S. Geol. Survey, 1972a; Richmond, 1977a). The area of abundant erratics at the site of the ice dam of Hayden lake shows an initial concentration there, but some erratics from this area were recycled westward to the Central and Solfatara Plateaus by full-glacial flow of Pinedale ice. The location of erratics south of the northern Yellowstone area (fig. 3) are from Howard (1937; fig. 15), Richmond and Waldrop (1975), Richmond (1977a), and Waldrop and Pierce (1975). Base from AMS 1:250,000 Ashton, Idaho, Mont., 1955.

Washburn Range is discussed in the section on Bull Lake Glaciation primarily because this is the age assigned to the lake sediments, which also require flow of ice from the Beartooth uplift. To my knowledge no relative age or stratigraphic evidence requires a late Bull Lake age rather than an earliest Pinedale age for either the lake sediments or for the emplacement of the erratics; emplacement during glacial buildup in early Pinedale time seems equally possible.

PINEDALE GLACIATION

The last major glaciation of the Yellowstone area, the Pinedale, left the most visible record; most surficial deposits in Yellowstone date from this glaciation. This section focuses primarily on the full-glacial conditions portrayed on plate 1*A*, and the interrelations between icemasses during glacial recession (pl. 2).

ICE ACCUMULATION PRIOR TO FULL-GLACIAL CONDITIONS

Little evidence has been obtained on the buildup of Pinedale ice in northern Yellowstone. According to glacial studies in the Great Lakes–St. Lawrence Lowland area and studies of the cores from the deep sea, global ice volumes may have built up by about 75,000 years ago and maintained at least 60 percent of maximum until deglaciation about 14,000 years ago (Gadd, 1971; Dreimanis and Karrow, 1972; Emiliani, 1966; Broecker and van Donk, 1970, p. 184; Shackleton and Opdyke, 1973, p. 45).

By analogy with this record, sizeable glaciers may have existed in Yellowstone in the interval from 75,000 to 14,000 years ago. The only radiocarbon dates from the Rocky Mountains relevant to this problem are from the southern part of Yellowstone Park (Birkeland and others, 1971, p. 219). Two dates on samples I collected there from lake sediments beneath Pinedale alluvial deposits yield infinite radiocarbon age dates (>38,000 (W-2012), >45,000 (W-2411)); Meyer Rubin, written commun., 1968; Richmond, 1970a; Birkeland and others, 1971, p. 219). Three dates on samples collected by G. M. Richmond from sediments immediately beneath either Pinedale Till or Pinedale kame deposits also have yielded infinite C-14 dates (>42,000 (W-2264), >42,000 (W-2197), >29,000 (W-2582); Meyer Rubin, written commun., 1969, 1970, 1971; Richmond, 1970a; Birkeland and others, 1971, p. 219, Richmond, 1972, p. 320). Sample W-2582 was reported to be 29,000 years old (Richmond, 1972, p. 320), but should have been reported to be >29,000 years (Meyer Rubin, written commun., 1973). No samples have yielded finite ages in the 15,000 to 35,000 year range. These age determinations suggest the Pinedale Glaciation started >40,000 years ago. Hydration rinds on obsidian from the Pinedale terminal moraines near West Yellowstone indicate an average age of glacial abrasion of about 30,000 years, and that some Pinedale glacial abrasion occurred more than 40,000 years ago (Pierce and others, 1976, fig. 4).

GLACIERS FROM THE BEARTOOTH UPLIFT

A glacier advancing down the Black Canyon of the Yellowstone River from the Beartooth uplift blocked and dammed the Gardner River just south of Gardiner, Mont. On the west side of MacMinn Bench (Pierce, 1973b, section 3), well-bedded lake sediments grade upward into contorted lake sediments containing lenses of till, which in turn grade into a thick section of till containing erratics of Precambrian rocks (probably from the Black Canyon of the Yellowstone); next in the series is till containing erratics, probably from the Gallatin Range. These relations suggest that ice from the Beartooth uplift dammed a lake and then overrode the sediments deposited in it before glaciers coming down the Gardner River from the Gallatin Range reached this area.

Several beds of angular coarse gravel immediately underlie the lake sediments. These gravel beds display openwork texture (lacking fine-grained matrix) and contain boulders with percussion spalls; they fill a pre-Pinedale valley of the Gardner River and apparently conformably underlie the lake silts. A prominent bed that contains abundant basalt boulders more than 1.5 m across can be traced upstream from the north end of MacMinn Bench for about 3 km to where it crops out beneath alluvial fans on the west side of Mount Everts and is clearly visible from the Mammoth-Gardiner road. These gravels probably indicate floods caused by sudden release of a succession of lakes similar in origin to the one in which the overlying lake sediments were deposited.

Ice from the Beartooth uplift advanced southward probably against and possibly past Specimen Ridge and the Washburn Range in early Pinedale time. Such an advance is postulated as the early glacial analog to the deglacial Deckard Flats readjustment, which will be discussed in a later section. During partial-glacial conditions, the Beartooth uplift appears to have been a more important ice source than the Yellowstone plateau.

Glaciers from the Beartooth uplift descended Soda Butte Creek and first blocked and then ascended the upper Lamar drainage, forming ice-dammed lakes. Several stratigraphic sections along the Lamar River and Cache Creek (Pierce, 1974b, sec. 8, 9, 10, 12) indicate ponding and deposition of sand and silt prior to

the arrival of Pinedale ice. Along South Cache Creek, till exposed in a thick section (Pierce, 1974b, sec. 8) is rich in tuff of the Pleistocene Yellowstone Group, which crops out only downstream; ice that deposited this till carried tuff up South Cache Creek before glaciers advancing down the creek reached this locality.

LOCAL ICECAP ON THE WASHBURN RANGE

A local icecap existed on the Washburn Range prior to full-glacial time. Most evidence for this icecap has been obliterated by flow during full-glacial conditions, except for striations on top of Observation Peak (southwesternmost peak of Washburn Range) indicating flow to the southwest, which are truncated by striations indicating flow to the north (fig. 15). Both sets of striations are equally fresh. The sequence is established by geometric relations, such as physical superposition. Also striations show southwest flow on the north sides of knobs bearing the northerly striations; this pattern indicates a sequence of southwesterly flow followed by northerly flow. This southwesterly flow is responsible for at least some of the Eocene volcanic erratics from the Washburn Range that are common on the rhyolite plateau just south of the Washburn Range.

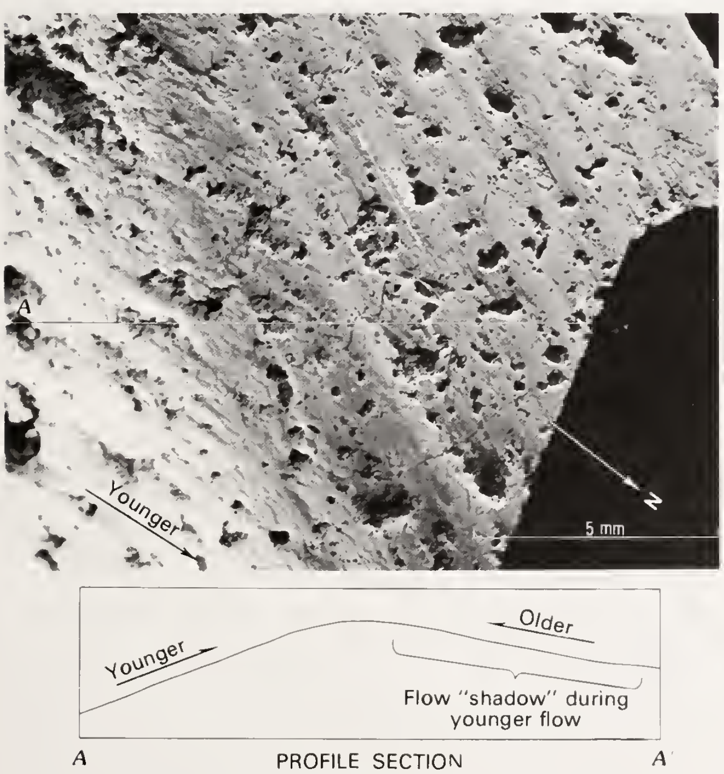


FIGURE 15.—Crossed striations on basalt of Observation Peak, altitude 2,864 m (9,397 ft). Both sets of striations are equally fresh and date from phases of the same glaciation. The younger striations enter into the edge of the area displaying the older striations but do not obliterate them, apparently because of the flow-shadow effect illustrated on the profile section (no scale).

ICECAPS ON THE YELLOWSTONE PLATEAU AND GALLATIN RANGE

A large icecap formed on the Yellowstone plateau in the southern part of the study area (pl. 1*A*). This icecap might have evolved from piedmont glaciers advancing onto the plateau from the adjacent ranges. However, it is also possible that permanent snow and ice began accumulating on the plateau surface itself. This is not unreasonable, for if the Pinedale glacial climate was 6°C colder than at present (Flint, 1971, p. 416; Heusser, 1965, p. 479), the plateau would have had a mean annual temperature of about −5°C (present MAT being approximately 1°C, fig. 4).

Once an icecap became established, it would tend to be self-amplifying: accumulation of ice would increase the elevation of the surface, which would increase orographic snowfall. An icecap on the Yellowstone plateau was in a very favorable position to intercept moisture from storms moving eastward up the Snake River Plain.

In the Gallatin Range, glacial buildup first produced mountain-valley glaciers of early Pinedale age prior to the formation of the full-glacial icecap there.

FULL-GLACIAL CONDITIONS

A reconstruction of the northern Yellowstone area under full-glacial conditions is shown on plate 1*A*, "full-glacial" here meaning when the main glaciers in the northern Yellowstone area were at or near their maximum volumes. During full-glacial conditions, most ice in northern Yellowstone National Park flowed toward terminal moraines in the Yellowstone valley. Although this report emphasizes the glaciation of northern Yellowstone National Park, it includes source areas and the main terminal area, which are outside the park. The terminal area of northern Yellowstone glacier is first described, followed by a description of the glacier up the valley of the Yellowstone to the park. Convergence of four wedge-shaped ice streams forming the accumulation area of the northern Yellowstone glacier is described and the wedge-shaped ice streams followed to their heads at ice divides. Finally, the full-glacial flow on the opposite side of these ice divides is discussed.

TERMINAL AREA OF THE NORTHERN YELLOWSTONE GLACIER

PINEDALE OUTWASH FAN AND ICE-MARGINAL CHANNEL

The Pinedale terminus of the northern Yellowstone outlet glacier lay 55 km north of the park boundary (pl. 1*A*), in an 8-km-wide part of the valley floor of the Yellowstone River. There melt water from the glacier formed two major fluvial features—a large outwash

FIGURE 16.—Outwash fan of the northern Yellowstone glacier displaying detailed, essentially unmodified, braided-channel morphology. Melt waters from a glacial system 3,400 km² in extent built this thick outwash fan. As shown by the dashed and dotted line, Pinedale ice reached as far north as the center of the photograph and stagnated, leaving ice-disintegration moraines. Photograph by U.S. Geological Survey, 1949.

fan and a deep ice-marginal channel. The Yellowstone River drained about 3,400 km² of glacial ice, and discharged large volumes of melt water.

The excellent preservation of braided-channel morphology on the large outwash fan (fig. 16) indicates that the fan is of Pinedale age. At its head, the fan surface is 45 m above the Yellowstone River; 10 km downvalley it is only 10 m above it; and 16 km downvalley it has merged with the present flood plain. Consequently the fan head owes its height above the river to the accumulation of outwash on the valley floor and the subsequent trenching of these unconsolidated deposits by the Yellowstone River. Therefore, the amount of incision of the outwash and end moraines does not necessarily indicate a pre-Pinedale age as inferred by Horberg (1940, p. 297), because the depth of incision here does not result from the same factors as the general increase in age with height above modern base level noted elsewhere in the Rocky Mountains.

On the northwest slopes of the Yellowstone valley, the edge of the outlet glacier is well delimited by a deep ice-marginal channel (pl. 1*A*). This channel has a gradient of about 35 m/km (Montagne, 1970, p. 15). Weed (1893, p. 27–30) observed this channel and similar channels farther downslope and described them as follows:

FIGURE 17.—Eightmile terminal moraines of Pinedale age with well-preserved glacial morphology. The area in the photograph beyond the Pinedale moraines (beyond dashed line) exhibits no evidence of a pre-Pinedale northern Yellowstone outlet glacier, although the terrain there is locally favorable for preservation of such evidence if it did exist. Oblique aerial photograph, looking east, by W. B. Hall.

Foothill canyons.—A most curiously interesting feature of the drift-covered area is the presence of transverse canyons cut in the breccias directly across the slopes of the foothills. Their constant relation to the ice margin suggests that they had their origin during the period of glacial occupancy of the valley. They are of considerable length, sometimes presenting several miles of unbroken walls. Their general direction is a most striking anomaly in the general structure and topography of the neighborhood, for the canyons cut across the slopes sometimes in apparent indifference to deeply trenched drains and rocky spurs alike * * *. Sometimes there are two or even three parallel canyons, in which case the highest is the largest and longest.

* * *the slopes west of the outermost canyon were bare of drift, and had not been covered by ice—this was positively established; in only one instance was any drift observed west of the cutting, and the boulders formed a strip but a few yards wide* * *. The slopes east of the canyon are heavily mantled with drift* * *.

Although Weed recognized the channels and deduced that the outermost one delimited the extent of glaciers from Yellowstone, he refrained from concluding that they were formed by ice-marginal streams. He did not think such ice-marginal streams would have flowed long enough to have cut 30 m into bedrock.

END MORAINES

Pinedale end moraines floor most of the Yellowstone valley for about 25 km upvalley from the outwash fan. An outer set of these moraines (fig. 17) forms a large horseshoe-shaped mass about 10 km across and 11 km

long (Weed, 1893; Horberg, 1940; Montagne, 1968, 1970), and are here called the Eightmile moraines, after Eightmile Creek, which crosses the northern margin of the moraines. An inner set lying 7 km upvalley are called the Chico moraines (Horberg, 1940; Montagne, 1970). Their outer margin lies along a major outwash channel in which Chico Hot Springs and the main road to the springs are located (pl. 2).

Although Horberg (1940) and Alden (1953, p. 185) referred the Eightmile moraines to the Bull Lake and the Chico moraines to the Pinedale, there is little basis for this large an age difference. As noted by John Montagne (1970) and Cliff Montagne (oral commun., 1970), soil profiles examined on both sets of moraines are similar and weakly developed; a B-horizon, if present, is weakly developed and less than 15 cm thick. Comparison of places where glacial morphology is best developed on both sets of moraines indicates equally strong and hummocky morphologic expression, with steep-sided, cone-shaped, closed depressions. The Chico moraines seem to contain more boulders and hence their surface is locally more bouldery.

Horberg (1940, p. 297) following Alden (1932, p. 123) separated the Yellowstone valley moraines into "early and late Wisconsin substages," which he correlated with the Bull Lake and Pinedale Glaciations, "largely upon the fact that the late Wisconsin terminal moraine descends into the inner post-early Wisconsin valley of the Yellowstone about 2 miles [3 km] north of Emigrant." But later statements by Alden (1953, p. 186) are indefinite on whether the Chico moraines descend into the inner valley. The bouldery ridges I examined in the inner valley are probably not moraines, but flood deposits. (See end of section on late Pinedale and younger floods.) The argument for a base level change between deposition of the two moraines does not hold water, for the flood of the main melt-water channel from the Chico moraines merges with an extensive outwash plain lying 45 m above the Yellowstone River, the *same* height above the river as the outwash fan from the Eightmile moraines.

A huge complex kettle in the Chico outwash plain extends more than a mile beyond the front of the Chico moraine (Montagne, 1968, p. 5). This kettle is more than 20 m deep, and formed by melting of the ice that advanced to the Eightmile moraines; but before ice melted to form the kettle, it was buried with outwash from ice at the Chico moraines. Therefore, the Chico and Eightmile moraines date from the same glaciation.

Although loess is generally thin or absent on Pinedale deposits, loess as much as 5 m thick occurs locally in drifts in the lee of some morainal ridges in the Chico moraines downwind (east) from the Yellowstone River.

Horberg's observations (1940, p. 298) also suggest that both the Eightmile and Chico moraines may be Pinedale. He implied that the evidence for differentiating the moraines of the west Mount Cowen glaciers into Bull Lake and Pinedale was much better than that for separating the moraines of northern Yellowstone outlet glacier. Alden (1932, p. 123) also expressed doubt that the Eightmile moraines are of Bull Lake age, and suggested the possibility that they might date from an early phase of the last glaciation.

Washing by abundant melt water appears to be responsible for the subdued morphology common on most of the Eightmile moraines. Locally, the morainal deposits show doughnut shapes (fig. 16) similar to those described by Gravenor (1955) and Clayton (1967, p. 32) as ice-stagnation features. Melt water was less abundant on the southeast side of the valley; hence, strong glacial topography (fig. 17) is locally well displayed in the form of sharp-crested morainal ridge and steep-sided depressions.

At Elbow Creek, the Eightmile moraine truncates local moraines correlated with the Pinedale (John Montagne, written commun., 1973).

NORTHERN YELLOWSTONE OUTLET GLACIER FROM ITS TERMINUS TO GARDINER

It is important to know the relative importance of ice sources contributing to the northern Yellowstone outlet glacier between the terminus and Gardiner compared with that from sources upstream from Gardiner. The evidence indicates that both under full-glacial conditions and during Pinedale recession, flow from tributary glaciers was minor or insignificant compared with outflow from sources upstream from Gardiner. (See also section entitled, "Phase relations between areas feeding the northern Yellowstone outlet glacier in pre-Deckard Flats time.") The importance of this point is that the relative size of the glacier in the Yellowstone valley is therefore directly related to the extent of glaciation in its primary source areas upstream from Gardiner. Establishing the primary source for the outlet glacier near Gardiner then becomes fundamental to resolution of a difference of opinion concerning the height of Pinedale ice in the source area of the northern Yellowstone glacier in Yellowstone National Park, where one published reconstruction (Richmond, 1969, fig. 2; U.S. Geol. Survey, 1972a) indicates the surface of Pinedale ice as much as 750 m lower than that defined by this study.

Between the outlet-glacier terminus and Gardiner, ice flow from about 10 tributary valleys that join the Yellowstone valley produced only minor effects on the

flow of the outlet glacier in the Yellowstone valley. Glacial erosion features, even on high interfluves between large tributary valleys, parallel the Yellowstone valley, indicating dominance of outlet glacier flow over that from the tributary glaciers (pl. 1*A*).

The Pinedale outlet glacier left well-preserved scour and depositional features along the valley walls where the topography and bedrock were favorable. Part of the outlet glacier flowed across Precambrian crystalline bedrock in the saddle high on the east side of Dome Mountain more than 600 m above the Yellowstone River. Scouring of the Precambrian bedrock there into whale-back forms (fig. 18) indicates vigorous ice flowage; the freshness of these features suggests a Pinedale age.

On the west side of Dome Mountain, Yankee Jim Canyon (pl. 1*A*) is the narrowest constriction in the glaciated part of Yellowstone valley. In 1881, Archibald Geikie described glacial features in Yankee Jim Canyon as follows (1881, p. 7–8):

> The rocky sides * * * are smoothly polished and striated from the bottom up apparently to the top. Some of the detached knobs * * * were as fresh in their ice polish as if the glacier had only recently retired from them. * * * There could be no doubt now that the Yellowstone glacier was massive enough to fill the second cañon to the brim. * * * The ice had perched blocks of granite and gneiss on the sides of volcanic hills not less than 1,600 feet [500 m] above the present plain of the river, and it not merely filled the main valley, but actually over-rode the bounding hills so as to pass into some of the adjacent valleys.

Between Yankee Jim Canyon and Gardiner much of the Yellowstone valley is either steep slopes underlain by soft rocks where landsliding and erosion have obliterated most glacial features, or bottom lands where late-glacial flooding and postglacial alluviation have covered the valley floor. But in gently sloping areas high on the valley walls, glacial features are well preserved (figs. 18–20). Weed (1893, p. 15–18, pl. 1*A*) noted abundant evidence of glaciation, including features supporting the conclusion that flow of the outlet glacier was dominant over local tributary glaciers. As shown on plate 2, erratics from Cinnabar Mountain were carried northwestward parallel to the Yellowstone valley and deposited at an elevation of 2,250 m (7,400 ft) near Cottonwood Creek (Weed, 1893, pl. 1, p. 17). This indicates that glaciers flowing down Mol Heron and Cinnabar Creeks did not cross the line defined by the source of the erratics and site of deposition, and that augmentation of the outlet glacier by local glaciers was of minor importance. (See p. F54). Glacial erosional features between Cinnabar Mountain and Cottonwood Creek also indicate flow parallel to the Yellowstone valley rather than from the tributary valleys.

FIGURE 18.—Glacial scour features on Dome Mountain divide, 600 m above the Yellowstone River. The northern Yellowstone outlet glacier filled Yankee Jim Canyon (pl. 1*A*) and spilled across the high pass in upper right and flowed towards left side of photograph. Note drumloid topography formed on scoured bedrock of Precambrian gneisses and Eocene volcanic rocks. Glacial polish is still apparent on roches moutonneés on the divide. Oblique aerial photograph by W. B. Hall.

FIGURE 19.—Erratics on glacial pavement high above the Yellowstone River near Gardiner, Mont. These fresh glacial features indicate that the ice was at least 1,100 m thick and that flow was parallel to the Yellowstone valley rather than outward from tributary valleys. Photograph by Cliff Montagne.

Five km north of Gardiner a deep melt-water channel (Pierce, 1973b) and fresh glacial pavement bearing scattered erratics (fig. 19) at an altitude of 2,700 m (8,800 ft) indicate Pinedale ice at that height. Directly across the Yellowstone valley on the north spur of Sepulcher Mountain, erratics, fresh striations, and ice-scoured bedrock (fig. 20*A*, *B*) show that Pinedale ice was at an altitude of 2,700 m (8,800 ft). On the north spur of nearby Electric Peak, fresh glacial scour indicates northwestward flow at an elevation of 2,800 m

FIGURE 20.—Glacial-scour features on the north spur of Sepulcher Mountain. *A*, Oblique aerial photograph showing glacial scouring of Eocene bedrock. Scour extends to the top of the grassy area along the ridge crest, 1,100 m above the Yellowstone valley floor. Arrow indicates ice flow from right to left. Photograph by W. B. Hall. X, site of *B*, Photomicrograph of striated and polished bedrock outcrop showing degree of preservation of striations and polish.

(9,200 ft), higher here probably due to the influence of local ice from cirques on Electric Peak.

The soft bedrock in the section of the Yellowstone valley between Yankee Jim Canyon and Gardiner was glacially scooped out, forming a bedrock basin upstream from the resistant bedrock threshold at Yankee Jim Canyon. Water wells in this section of the valley are still in fine-grained sediment (R. Miller, oral commun., 1972) at depths 15 m below the level of bedrock farther downvalley in Yankee Jim Canyon.

From Gardiner to the terminal moraines, the northern Yellowstone outlet glacier descended at the gradient shown on plates 1*A* and 4, as determined by observation of erratics and glacial scour features.

ICECAP SOURCES FEEDING THE NORTHERN YELLOWSTONE OUTLET GLACIER

During full-glacial conditions, ice from four main sources converged toward Gardiner to form the northern Yellowstone outlet glacier (pl. 1*A*). The converging ice streams are differentiated somewhat arbitrarily inasmuch as they formed a continuous icemass. These streams are defined by directions of scour features and transport of erratics under full-glacial conditions and are designated as the Gallatin Range icecap, the plateau icecap between the Gallatin and Washburn Ranges, the plateau icecap overriding the Washburn Range-Specimen Ridge area, and the Beartooth-uplift icecap. The northern Yellowstone outlet glacier and the ice from these icecaps that fed it together constituted the northern Yellowstone glacier.

Figure 21 illustrates the convergence of ice streams to form the northern Yellowstone outlet glacier.

GALLATIN RANGE ICECAP

The westernmost of the four wedge-shaped icestreams headed in an icecap on the Gallatin Range, flowed northeast along the Gallatin Range front and streamed around Sepulcher Mountain (pl. 1*A*). The glacial streaming, molding, and scouring features produced by this flow are well displayed in Gardners Hole (frontispiece, fig. 22), where parallel drumloid ridges cover some 40 km^2 of meadowland. Two features indicate that the direction of flow was northeasterly rather than southwesterly along the drumloid trend (fig. 22). First, the trend of drumloid features is oblique to the mountain front (frontispiece, pl. 1*A*). Ice at the base of a glacier would not have formed such features if it flowed into the front, but could have formed such features by flowing down and obliquely away from the front. Second, as noted in figure 13, erratics from the south end of the Gallatin Range were carried in a northerly direction parallel to the drumloid terrain.

Fresh glacial-scour features delimit the approximate altitude of the Pinedale ice-surface at the north end of the Gallatin Range ice-stream (pl. 1*A*). These features (fig. 20) extend up to an altitude of 2,700 m (8,800 ft) on the north side of Sepulcher Mountain and to about 2,750 m (9,000 ft) on the south side (pl. 1*A*). No evidence was found to indicate that the higher parts of Sepulcher Mountain (altitude 2,942 m or 9,652 ft) were ever glaciated. Southeast of Sepulcher, ice-scoured travertine and erratics from the Gallatin Range are scattered over Terrace Mountain (altitude 2,442 m or 8,011 ft). Southwest of Sepulcher Mountain, at altitides of about 2,900 m (9,500 ft), the upland flats of Quadrant and Little Quadrant Mountains are glacially scoured and smoothed, but no erratics are present; ap-

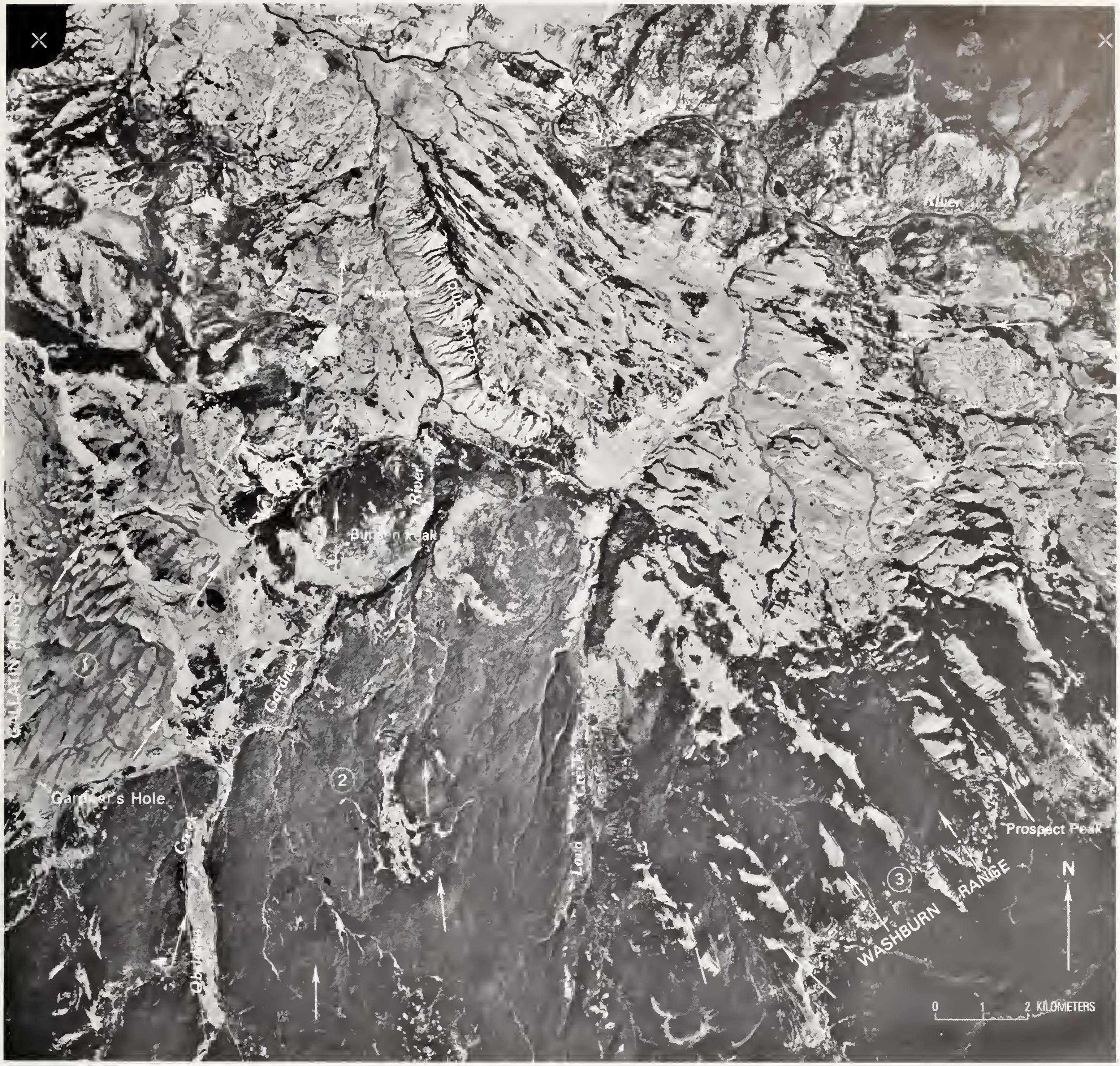


FIGURE 21.—Convergence of ice streams towards Gardiner to form the northern Yellowstone outlet glacier. 1, Drumloid topography in Gardners Hole (also frontispiece) and adjacent areas formed by northeasterly to northerly flow from the Gallatin Range icecap and the plateau icecap between the Gallatin and Washburn Ranges. 2, Northerly flow between Obsidian and Lava Creeks left striations and basalt erratics on Bunsen Peak. 3, Striations and scour features cross the crest of the northwestern part of the Washburn Range. 4, Glacial-scour features on Mount Everts and adjacent uplands indicate flow swinging from westerly to northwesterly. Dot in arrow indicates striations observed on the ground. High-altitude aerial photograph by NASA, 1969.

parently flow was outward from, rather than toward these mountains. These observations indicate that the ice surface was above 2,900–3,000 m (9,500–10,000 ft) in the Gallatin Range. Consequently, ice east of the front of the Gallatin Range was 600-700 m thick.

A variety of other evidence supports the conclusion that ice east of the Gallatin Range front was hundreds of meters thick. First a ridge of Eocene volcanic rocks trends northwest across Gardners Hole. Although the ridge projects 90–180 m above Gardners Hole, scour

FIGURE 22.—Oblique view northwest over drumloid topography in Gardners Hole. Flow direction is from left to right. A weakly developed soil of Pinedale age is developed on the surface till, but a clayey reddish weathering zone on an underlying till of inferred Bull Lake age is exposed where the Gardiner River has cut into the core of some drumlins. Photograph by W. B. Hall.

and flow features (frontispiece, fig. 22, pl. 1*A*) trend undeflected across it. Second, drumloid features across the mouths of valleys at the Gallatin Range front (especially Panther and Indian Creeks, and an unnamed creek north of Winter Creek) indicate that the ice on the flats to the east was as high as that in the valleys of the Gallatin Range, where well-preserved evidence of glaciation extends more than 600 m above the valley floors. Third, the drumloid features themselves indicate formation by ice probably more than 500 m thick.

PLATEAU ICECAP BETWEEN THE GALLATIN AND WASHBURN RANGES

A wedge-shaped ice-stream 600–900 m thick flowed northward from the plateau between the Gallatin and Washburn Ranges (pl. 1*A*). Near its north end the icecap overtopped Bunsen Peak (altitude 2,610 m; 8,564 ft); abundant basalt erratics and coarse glacial striations there indicate northward flow. On the upper part of Bunsen Peak, erratics of Gallatin Range rocks (andesitic intrusives, gneiss, quartz, limestone) are either absent or very much less abundant than they are below an altitude of about 2,300 m (7,500 ft). Although hundreds of basalt erratics, some well polished and striated, were noted above 2,450 m (8,000 ft) on Bunsen Peak, only one erratic of intrusive rock from the Gallatin Range was found, indicating that full-glacial flow was not easterly from the Gallatin Range. The basalt erratics were primarily derived from the Pleistocene Swan Lake Flat Basalt which underlies ex-

tensive areas to the south, about 300 m below Bunsen Peak (pl. 1*B*; U.S. Geol. Survey, 1972b). Thicknesses of weathered rinds on 13 surface erratics of sugary-textured basalt suggest a Pinedale age: five had weathered rinds 0–0.25 mm thick on glacially abraded surfaces and five had rinds 0.25–0.5 mm thick. The basalt erratics indicate that northward flow was vigorous enough to lift erratics 300 m above the altitude of their origin (fig. 27). These features require an ice mass more than 400 m thick just south of Bunsen Peak; northward flow features show that the ice surface continued to rise for about 16 km south of Bunsen Peak (fig. 27).

PLATEAU ICECAP OVERRIDING THE WASHBURN RANGE AND SPECIMEN RIDGE

Striated bedrock at more than 100 localities along the crest of the Washburn Range and on Specimen Ridge reveals that the last glacial flow across these mountains was towards Gardiner (pl. 1*A*). These striations occur on the highest parts of the ridge crest, which is locally above 3,000 m (10,000 ft) in altitude and generally above 2,700 m (9,000 ft). Prior to this study, evidence for regional glaciation had not been found as high as 2,700 m (9,000 ft) in ths area. At about 2,530 m (8,300 ft), Alden (1928) observed a glaciated rock surface with rat tails indicating flow to the north. A. D. Howard noted polished and striated bedrock as high as 2,690 m (8,825 ft).

Glacial striations and polish discovered on the crest of the Washburn Range in 1967 were first thought to be of Pinedale age. But Richmond (1969, fig. 2) concluded that the upper limit of Pinedale ice was near the base of the range (at the position mapped on plate 2 as the Deckard Flats limit), and it was determined therefore, that the glaciation of the crest was pre-Pinedale (Pierce, 1968). Further mapping and study of the glacial system forming the northern Yellowstone outlet glacier (pl. 1*A*) indicated to me, however, that a Pinedale age was more likely. Because this difference of opinion is still unresolved (U.S. Geological Survey, 1972a), considerable detail is given in this section.

The form of the Washburn Range (fig. 23) suggests that ice overrode the entire range. Its smooth and rounded appearance contrasts markedly with other mountainous terrain underlain by similar volcanic rocks of the Absaroka Supergroup. H. J. Prostka mapped the Absaroka rocks in the northern Yellowstone area and also noted this difference; he concluded the morphologic distinctness of the Washburn Range is not due to its bedrock geology (oral commun., 1973). Remarkably little surficial debris exists on much of the *upland* terrain; my observations as well as those of W. B. Hall (written commun., 1973) and H. J. Prostka (oral commun., 1973) suggest that bedrock is commonly present within 0.3 m of the surface.

The evidence of Pinedale full-glacial conditions in the Washburn Range consists primarily of scour features, for ice accumulation and abrasion prevailed at altitudes above 2,700 m (9,000 ft).

Flow across the crests of the Washburn Range and Specimen Ridge was northward towards Gardiner and not southward from the Beartooth uplift (fig. 24). The most commonly observed feature demonstrating this flow direction is the concentration of glacial striations and polish on the upflow side of bedrock protuberances (figs. 6, 7). In addition, low-angle friction cracks (fig. 6) consistently dip to the north. Where risers face north, striations and polish occur on treads but not on risers, showing northward flow (figs. 6, 7). Rat tails on conglomerate between altitudes of 2,530 and 2,740 m (8,300 and 9,000 ft) in the eastern part of the Washburn Range near Dunraven Pass clearly demonstrate ice flow to the north. The rat tails and striations do not trend down the west slope of Mount Washburn but are subhorizontal and indicate a thick glacier flowing and shearing horizontally along the side of the mountain.

Most of the striations occur on ridge crests. Therefore the flow that produced them was not affected by higher local topographic features. Consequently, these striations indicate relatively precisely the direction of flow and the surface slope of the ice.

Although most of the Absaroka volcanic rocks are so friable that they do not generally display striations, striations do locally occur on individual clasts of dense, very dark gray andesite in the volcaniclastic bedrock. These striations and the associated polish are very fresh (figs. 25 and 26). Photomicrographs (fig. 7) show that the striations are very fine, linear, parallel scratches, some less than 0.1 mm across. The rock surface shows some postglacial pitting where chemical or mechanical weathering, or both, has removed parts of individual rock grains, but on polished surfaces little or no weathering has occurred since glaciation. When the rock is broken perpendicular to the polished surfaces, usually no rind can be observed with the naked eye, indicating that a rind, if present, is generally less than 0.25 mm thick. Thin sections perpendicular to the striated surface show no clear rind, but a slight lightening of the groundmass and reddish staining of some ferromagnesian minerals occur to a depth of 0.5 mm.

Figure 27 shows that the lower part of the icemass south of the Washburn Range was nearly static. No erratics of rhyolite from the plateau were found high on the southern Washburn Range, and scour features are mostly restricted to an altitude above 2,700 m (9,000

FIGURE 23.—Oblique view, looking northward, of the southeastern part of the Washburn Range, showing its smoothed and rounded appearance. This form is thought to reflect overriding of the entire range by a Pinedale icecap whose axis lay just south of the range (pl. 1*A*). Bedrock is commonly present within 0.3 m or so of the surface at altitudes above 2,680 m (8,800 ft), and fresh striations indicating glacial flow to the north occur all along the range crest. Dunraven Pass has a glacially rounded, U-shaped form, and striations on the north side of the pass indicate flow northward through the pass. Photograph by W. B. Hall.

ft). Thick colluvium locally occurs on the southern slopes of the Washburn Range, also suggesting lack of glacial scour. This lack of scour and transport is attributed to: (1) the height of the range being considerably greater than where such shearing and transport have been noted elsewhere in the park (Bunsen Peak, Gardners Hole ridge) and (2) the axis of the icecap being probably only a few kilometers south of the Washburn Range, thus requiring too great a curvature of shear planes.

Amethyst Mountain (altitude 2,930 m or 9,614 ft) at the crest of Specimen Ridge displays striations and glacial polish (fig. 28), which indicate northwesterly flow during full-glacial conditions (fig. 29; pl. 1*A*). Such a flow direction here is particularly important because it indicates that during Pinedale full-glacial time, the plateau icecap was at the same height as ice from the Beartooth uplift. The resultant direction of flow between these two abutting icemasses was northwestward along their common boundary and toward the outlet glacier at Gardiner.

BEARTOOTH UPLIFT ICECAP

The largest and most obvious ice stream that contributed to the northern Yellowstone outlet glacier was

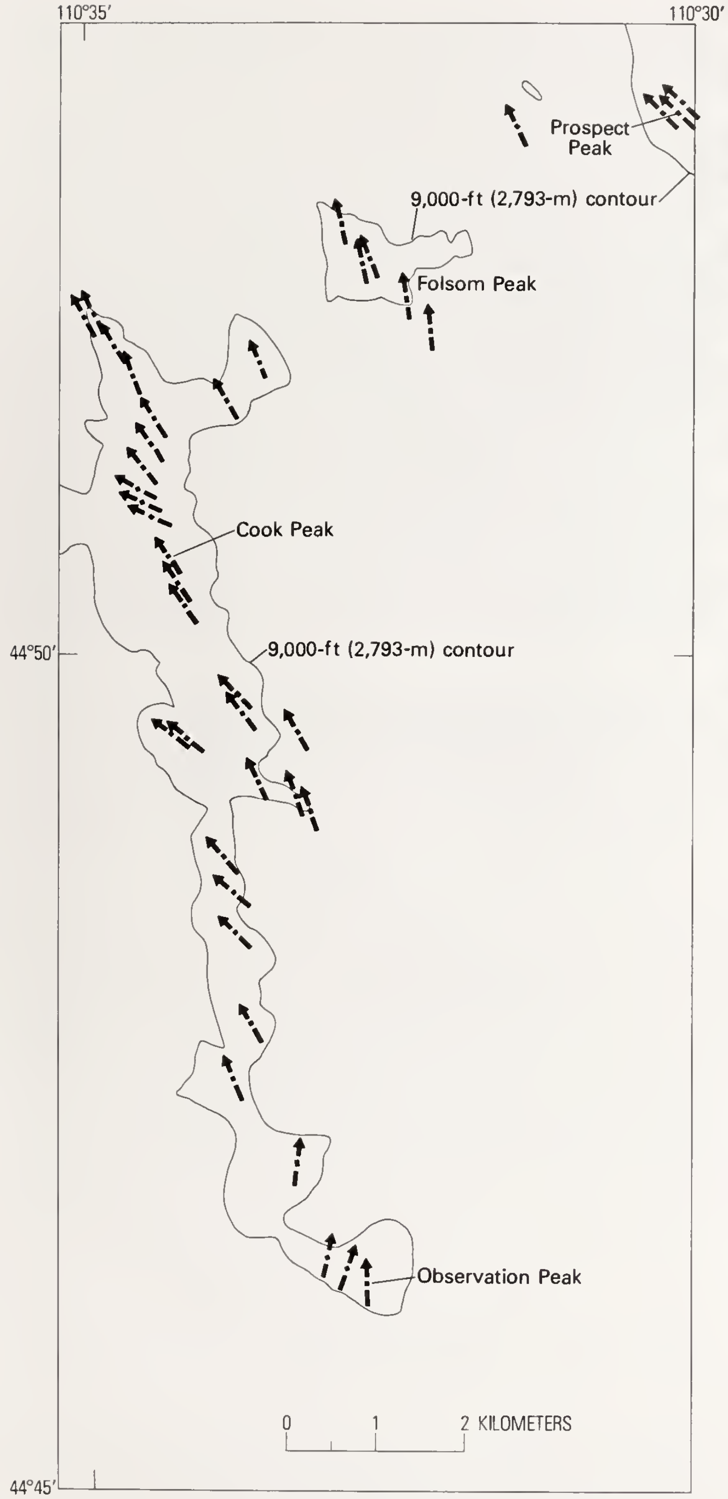


FIGURE 24.—Orientation of striations and direction of ice movement in the western part of the Washburn Range. All the 41 striation localities found above an altitude of 2,700 m (8,800 ft) are shown. Some systematic variation occurs along the range crest, but local variations are small. Dot, locality; line, striation orientation; arrowhead, flow direction.

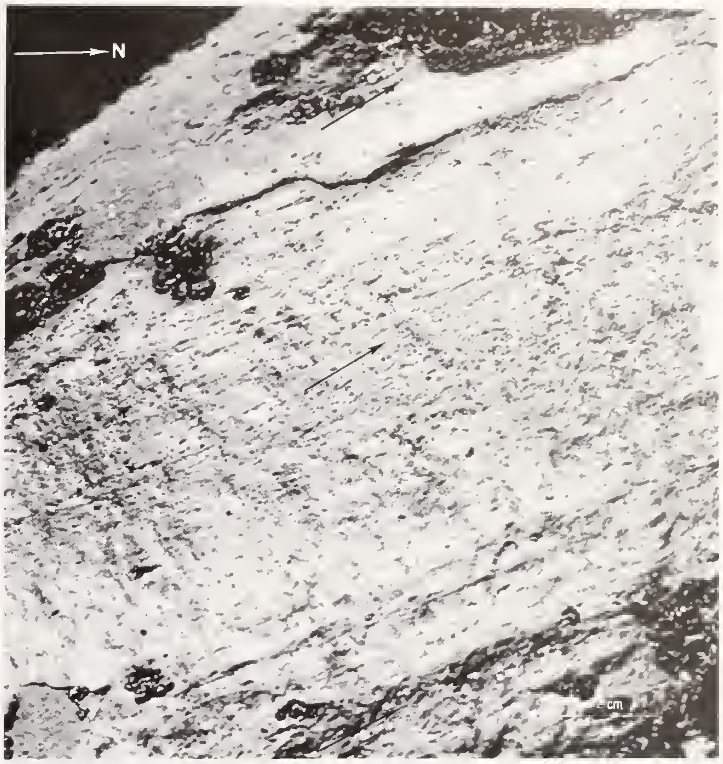


FIGURE 25.—Glacial polish and striations formed on vertical face at Observation Peak, southwest corner of the Washburn Range. Arrows indicate direction of full-glacial flow northward up and over this peak. Bedrock is Quaternary basalt.

from a large icecap on the Beartooth uplift. This stream flowed southwestward to the Lamar-Yellowstone drainage, then northwestward toward the outlet glacier.

On Palmer Mountain near Gardiner, at the downflow end of this mapped ice stream (pl. 1*A*), the altitude of the ice surface was about 2,770 m (9,100 ft). The upper several hundred feet of Palmer Mountain (altitude 2,853 m or 9,359 ft) consists mostly of frost rubble, and no evidence of glaciation was found. South of the mountain crest, evidence of glacial scour is common on the bench at an altitude of 2,700 m (8,800 ft). Ice-flow lineaments on this bench parallel the Yellowstone valley and are at a high angle to the drainage from local cirques.

Northeast of Palmer Mountain, towards the icecap on the Beartooth uplift, fresh glacial striations and scoured bedrock on uplands and ridgecrests indicate that the ice surface was above an altitude of 2,900 m (9,600 ft).

The upper limit of ice under full-glacial conditions is difficult to determine east of Palmer Mountain,

FIGURE 26.—Striations and glacial polish on breccia clast in Eocene bedrock at Prospect Peak (altitude 2,930 m or 9,525 ft) in the northern part of the Washburn Range. Note how striations and polish round upper edge of boulder and then disappear. About 10 similarly striated clasts protruding from the bedrock were found on the peak, each polished and striated on the southeast side, indicating flow to the northwest (to the upper right). Photograph by Cliff Montagne.

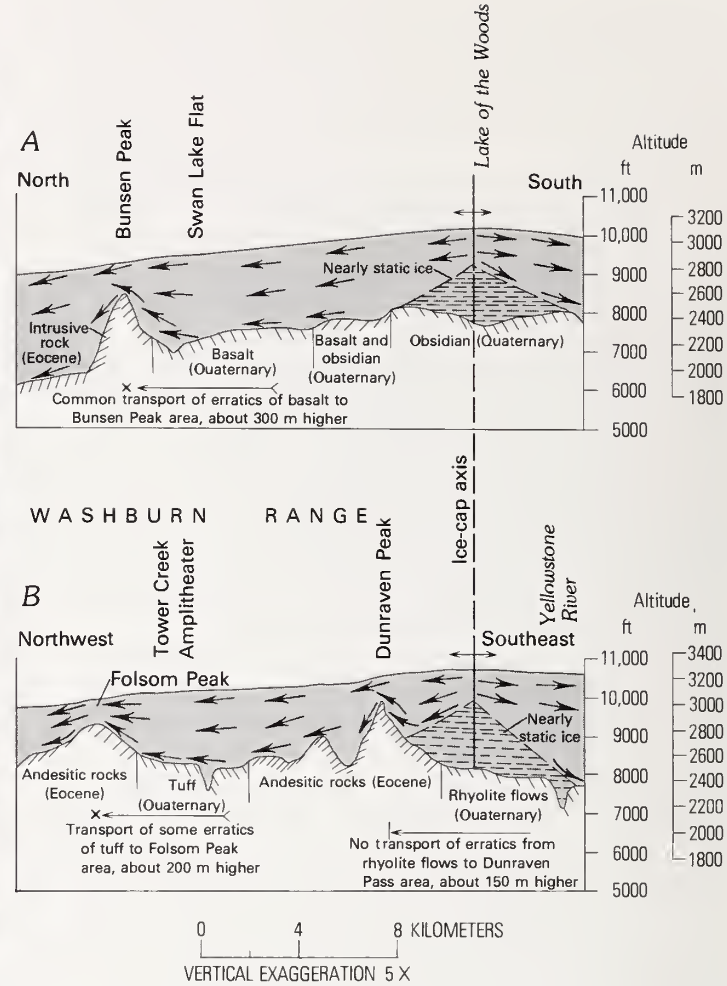


FIGURE 27.—Glacial flow across topographic obstacles. *A*, Flow over Bunsen Peak. Basalt boulders were carried upward about 300 m by ice from Swan Lake Flat to the top of Bunsen Peak (both south of Gardiner, pl. 1). No obsidian was found there, perhaps because the area of obsidian outcrop was overlain by nearly static ice along the axis of the icecap. *B*, Flow across the Washburn Range. Because of the height of the range and its proximity to the icecap axis, erratics of flow rhyolite were not carried high onto the southern part of the range. But farther north in the Folsom Peak area, which was neither as high nor as near the icecap axis, erratics of tuff from the Tower Creek area were lifted about 200 m above their place of origin.

because all of the high ridges and peaks for the next 50 km are intensely glaciated to their summits of 2,400–3,000 m (8,000–10,000 ft); the highest of these is Mount Hornaday, 37 km upvalley, at an altitude of 3,059 m (10,036 ft).

Within this 50-km interval, pervasive scour features indicate that an icemass flowed southwestward into the trench occupied by the Lamar-Yellowstone drainage. This icemass had a relatively flat upper surface but an irregular lower surface that conformed to the high relief of the present terrain. Its main flow-channels were the valleys of Hellroaring Creek, Buffalo Creek, Slough Creek, and Soda Butte Creek (pl. 1*A*). Figure 30 illustrates southward flow from the Hellroaring Creek drainage into the large westward-moving glacier in the Lamar-Yellowstone drainage.

Some of the highest mountain tops in the northeast corner of Yellowstone National Park display nearly flat terrain mantled with thick colluvium and show no evidence of glacial scour. On these small uplands above the ELA, snow apparently was blown away. The lowest altitude of these flat, apparently unglaciated upland surfaces are as follows: Abiathar Peak, 3,260 m (10,700 ft); Amphitheater Peak, 3,230 m (10,600 ft); and Cutoff Mountain, 3,200 m (10,500 ft). Below these uplands are cliffs of eroding bedrock where postglacial erosion has obliterated the upper limit of the Pinedale Glaciation.

An ice stream about 900 m thick and 6 km wide, heading near lower Aero Lake just outside the north-

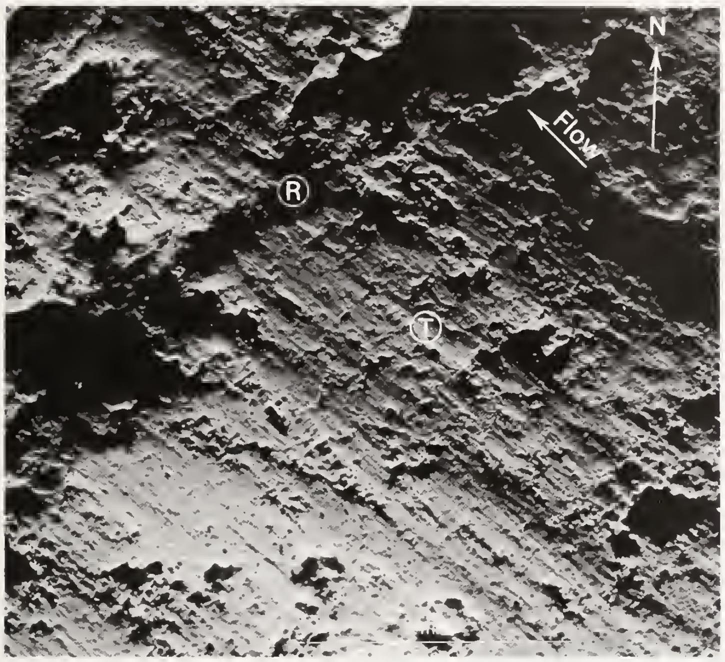


FIGURE 28.—Photomicrograph of striated and polished surface of Eocene bedrock on top of Amethyst Mountain (altitude 2,930 m; 9,614 ft), the highest peak on Specimen Ridge. Flow to northwest is indicated by arrow. Note tailing out of striations in direction of flow, and polish on tread (T) and not on riser (R) across stepdown in bedrock surface to the northwest. Similar striations and polish were found on about 10 boulders in the volcaniclastic bedrock.

east corner of the park, entered the Yellowstone drainage from the headwaters of Clarks Fork and, joining local ice, flowed down the valley of Soda Butte Creek. Thus, a continuous glacier 146 km long (pls. 1*A*, 4) extended from the head of Clarks Fork to the glacier terminus at the Eightmile moraines. This was probably the longest glacier in the United States, excluding those originating from the Canadian or Alaskan icecaps.

ICE IN THE UPPER LAMAR DRAINAGE BASIN

Evidence of the extent of ice in the upper Lamar drainage basin is most convincingly displayed along its headwaters along the east park boundary. There at more than 20 localities, glacial scour features, including striations, occur as high as 3,050 m (10,000 ft), and indicate flow eastward across the drainage divide. Downstream from the mouth of the upper Lamar, glacial scour features indicate that ice levels were equally high. Within the upper Lamar drainage basin, uncommon glacial flow features (pl. 1*A*; Pierce, 1974b) and the rounded character of the terrain also suggest that ice filled the drainage basin to altitudes above 3,050 m (10,000 ft).

Interpretation of the glaciation of the upper Lamar is complicated by the differences in flow between early

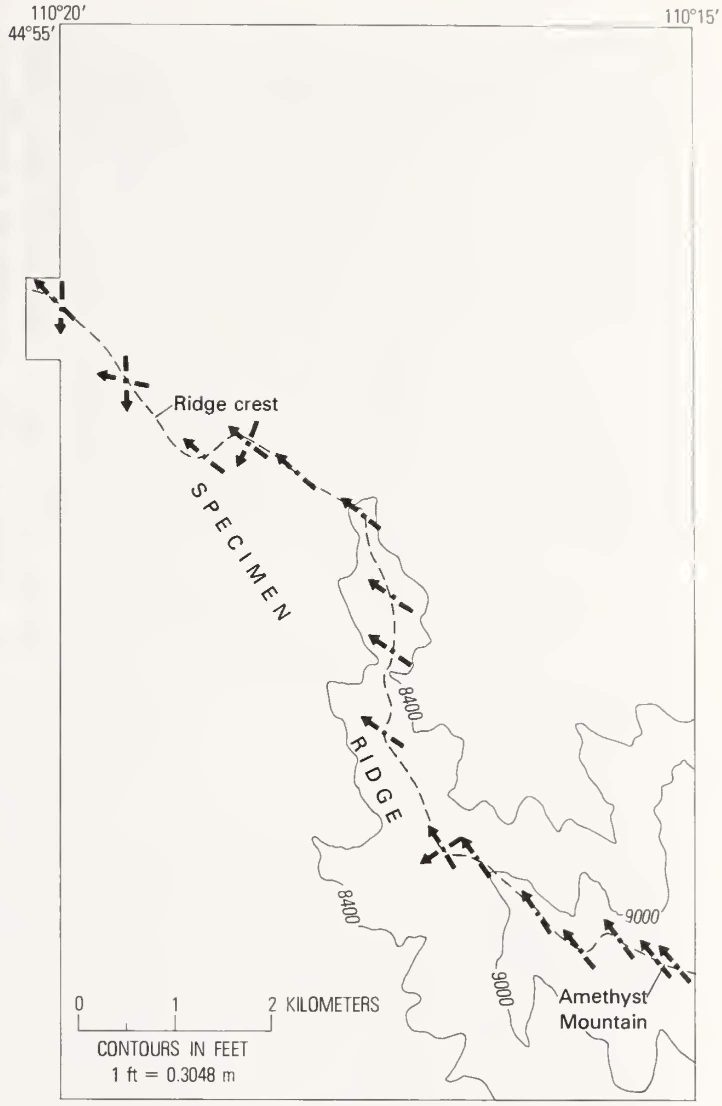


FIGURE 29.—Orientation of striations (arrows) along the crest of Specimen Ridge. Dots are localities of observation. Above an altitude of 2,750 m (9,000 ft) directional indicators (figs. 5, 28) indicate flow to the northwest. Below the 9,000-foot contour, geometric relations indicate that southerly striations are younger than the northwesterly ones. Also, erratics of Paleozoic limestone and Precambrian rocks below the 9,000-foot contour indicate an incursion of ice from the north, which occurred during the Deckard Flats readjustment. Base from U.S. Geological Survey 1:62,500 Tower Junction, 1959.

glacial, full-glacial, and late glacial times. Both before and after full-glacial time, ice from the Beartooth uplift flowed up the Lamar, formed ice-dammed lakes, and deposited till and erratics of Precambrian and Paleozoic rocks on the Eocene volcanic bedrock. The distribution of erratics from the Beartooth uplift (pl. 1*A*) indicated that Beartooth ice advanced only into the lower parts of the upper Lamar drainage. If ice from the Beartooth uplift had advanced up the Lamar and over the divide along the headwaters, then erratics from the Beartooth uplift should be widely distributed

110°30′

45°00′

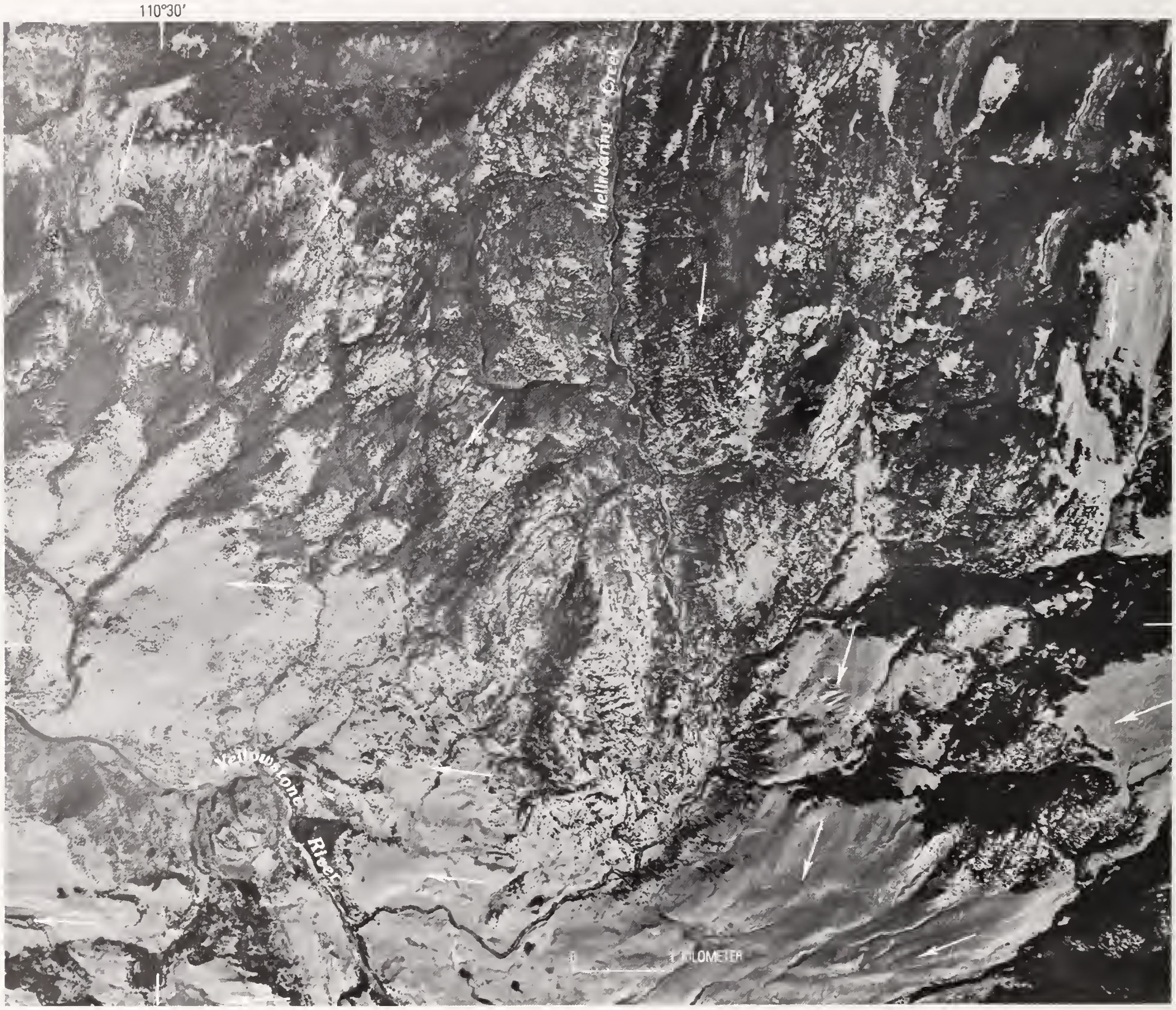


FIGURE 30.—Aerial view of glacial scouring, streaming, and molding features produced by ice from the Beartooth uplift as it flowed southward into the Yellowstone valley. Note strong convergence of flow with other ice coming down the Lamar-Yellowstone drainage. Photograph by NASA, 1969.

throughout the Lamar; they are not. Thus during full-glacial time, the upper Lamar was filled with ice as much as 900 m thick that accumulated there and flowed both eastward away from, and westward into, the northern Yellowstone glacier.

The absence of good evidence for glacial scour of most of the upper Lamar terrain may be explained by three factors. First, the Eocene volcanic bedrock in the upper Lamar generally displays evidence of glaciation poorly. Glacial scour is also uncommon on similar bedrock along the valley of Soda Butte Creek; this valley was occupied by a glacier more than 1,100 m thick coming from the Beartooth uplift. Second, the thicker and more active glaciers from the Beartooth uplift impeded outflow from the upper Lamar, resulting in thickening of the ice there. This thick ice probably flowed more by internal deformation than by basal slip, thus leaving little evidence of glacial scour. Third, this thickening also resulted in overflow of the divide along the east park boundary, thus displacing the ice-divide westward from the drainage divide (pl. 1*A*). East of this ice divide, scour features are to be expected only along the drainage divide. West of this divide, scour features may not occur within several kilometers of the icecap axis (fig. 27).

Although evidence of glacial scour (basal sliding) is poor, mass-balance considerations suggest that outflow from the upper Lamar area should have been

of the same magnitude per unit area as from other icecaps in northern Yellowstone. The average altitude of the ice surface in the upper Lamar was at least 3,100 m, about 300 m above the estimated equilibrium-line altitude in northern Yellowstone. Substantial accumulation must have occurred at this altitude, and total outflow to both the east and west would be a function of this accumulation multiplied by the surface area of the icecap in the upper Lamar (fig. 49).

The part of the northern Yellowstone glacier consisting of ice from the upper Lamar could not be traced by field observations of glacial scour or glacial erratics downstream from the upper Lamar. Downstream from the confluence of the upper Lamar with Soda Butte Creek, ice from the upper Lamar is for convenience included with ice from the Beartooth uplift (pl. 1*A*).

AGE OF THE HIGH-LEVEL SCOUR FEATURES

A reconstruction of the Pinedale Glaciation of Yellowstone indicates that most of the high-level glacial-scour features of the Washburn Range and Specimen Ridge are above the limit of Pinedale ice (Richmond, 1969, fig. 2; U.S. Geological Survey, 1972a). This reconstruction shows the altitude of the Pinedale limit at 2,740 m (9,000 ft) on Specimen Ridge, 2,560 m (8,400 ft) on the northwest side of the Washburn Range, and as low as 2,440 m (8,000 ft) on the south side of the Washburn Range. Unfortunately, the basis for these pre-Pinedale age assignments is not published.

Within the source area of the northern Yellowstone outlet glacier, the following points summarize my principal reasons for assignment of all the observed, well-preserved glacial-scour features to the last major, or Pinedale, glaciation:

1. The largest northern Yellowstone outlet glacier to deposit end moraines that retain morainal topography was of Pinedale age. Therefore Pinedale ice levels in the source area of this outlet glacier should have been as high as any previous glaciation that might have left fresh glacial-scour features in the source area. If the upper 300–600 m of fresh glacial features were formed by pre-Pinedale northward flow, then this glacier would have been much thicker, but shorter than the Pinedale.

2. The ice-surface profiles shown on plates 1*A* and 4 between the Pinedale terminal moraines and the high-level scour features on the Washburn Range and Specimen Ridge are consistent with the known relations between ice-surface slope and ice thickness. (See basal shear stress discussion and fig. 48.) Differing reconstructions of the extent of Pinedale ice (U.S. Geological Survey, 1972a) yield unreasonably low basal shear stress values. (See section "Basal shear stress and alternate glacial reconstructions.")

3. The relation between the orientation of scour features and contours on the ice surface requires the ice surface to be at similar altitudes in adjacent ice streams (pl. 1*A*). Mapping these directional flow features and contouring of the ice surface (pl. 1*A*) shows that the upper ice limit in the Gallatin Range, Beartooth uplift, Washburn Range, and uplands near Gardiner are all the same age.

4. No relative-age boundaries of major significance are apparent from the valley floors to the highest glacial features. For example, all the striations on dark-gray andesites and basalts along the crest of the Washburn Range and on valley floors are equally fresh, and most weathering rinds are less than 0.5 mm thick, indicating only one general age.

5. Most striations on rocks of the Washburn Range and Specimen Ridge are on 0.3–1-m boulders in the volcanic bedrock. These striated boulders are equally well preserved both above and below the "Pinedale limit" of the reconstruction of Richmond (1969, fig. 2) and U.S. Geological Survey (1972a). Since they were formed, the matrix around these clasts has weathered so that most could be readily dislodged, regardless of whether they are high or low in the terrain. In places, striations from the highest glacial phase are crossed by those of a lower glacial phase (Pierce, 1973b, 1974a). If the glacial scouring that formed the two sets of striations were separated by a nonglacial weathering interval, it seems likely that weathering and frost action during such an interval would have loosened the boulders so that the later glaciation would have dislodged them.

6. Soil-profile development since deglaciation does not indicate that the high glacial terrain is significantly older than the last glaciation. As summarized in Pierce (1974b), profiles related to the highest levels (T and M) are weakly developed. Similar development of profiles is found above and below the Deckard Flats boundary. (Deckard Flats is the late phase of the recessional Pinedale, discussed later.) Some profiles below the Deckard Flats limit are deep and show clay films (Pierce, 1973b, profile L). All soil profiles possibly relating to a high-level glaciation are within the limits of variability of profiles at lower levels and thus do not indicate any significant age difference.

7. Ice levels and ice thicknesses great enough to overtop the Washburn Range are not unreasonable, for there is agreement that only 30 km south of the Washburn Range in the Yellowstone Lake area a large Pinedale icecap reached an altitude of 3,200 m (10,500 ft) (Richmond and Pierce, 1968; Richmond, 1969). The main difference of opinion is whether or not the

Pinedale ice surface decreased by more than 600 m between the Yellowstone Lake area and the Washburn Range (Richmond, 1969), or maintained an altitude of about 3,200 m in a northerly direction along its axis. In addition to arguments based on field relations, the 600 m decrease in ice surface altitude seems excessive, for such a drop would result in basal shear stresses greater than 2.5 bars, well above general range of 0.5–1.5 bars observed on glaciers. (See p. F67.)

ICE DIVIDES AND FLOW AWAY FROM THE NORTHERN YELLOWSTONE GLACIER

Ice divides existed on the icecaps at the heads of the four ice streams just described. Most of these ice divides were not coincident with drainage divides; hence ice flowed across almost all of the present drainage divides. The icemasses on the far side of icecap divides are complementary to the northern Yellowstone glacier, and a reconstruction of the upper limits of the Pinedale icemass (pl. 1*A*) must be conformable across these divides.

GALLATIN RANGE

Under full-glacial conditions, ice west of the icecap axis flowed westward across passes along the crest of the Gallatin Range and around the south end of the range. This flow augmented local glaciers, which advanced about 15 km downvalley from the drainage divides of the Gallatin River and Grayling Creek. Westward-flowing ice about 300 m thick scoured seven passes (pl. 1*A*; fig. 31), most of which are at altitudes just above 2,750 m (9,000 ft.)

The terminal moraines of the glacier in the Gallatin valley lie in a gently sloping area between the Gallatin and Madison Ranges (pl. 1*A*). The moraine pattern in the terminal area is complex and demonstrates that flow occurred across local drainage divides. (See pl. 1*A* and fig. 37.) As discussed later (p. F51), the moraine crests inside the terminal moraine of the Gallatin River valley glacier form an important Pinedale recessional sequence.

Along Grayling Creek (fig. 37) about 2.4 km upstream from U.S. Highway 191, large multicrested, bouldery moraines mark the margin of Pinedale ice. The moraines are as bouldery and sharp crested as Pinedale moraines formed from granitic debris elsewhere in the Rocky Mountains. Their well-developed constructional topography and boulderiness result from large blocks of Precambrian gneiss supplied from an area aptly named "The Crags." The glacier in this valley was 450 m thick and was much too large to have been derived solely from several small low-altitude cirques in The Crags and one poorly defined cirque at the head of the valley. However, several glacial scoured divides at the head of Grayling Creek reveal one additional source of ice (pl. 1*A*). However, this addition was probably not as large as that from around the south end of the range. On the upland adjacent to the poorly defined cirque between the two forks of Grayling Creek, the bedrock is glacially rounded and scoured, and striations and scour features indicate northwestward flow towards Grayling Creek.

GLACIATION SOUTH OF THE GALLATIN RANGE

The axis of the plateau icecap lay just north of the southeast end of the Gallatin Range (pl. 1*A*). Beneath the ice divide, flow features are not apparent because the immediately overlying ice was nearly static (fig. 27). Several kilometers to the southwest, however, abundant glacial-scour features and erratics demonstrate that flow was to the southwest (pl. 1*A*; fig. 32). This flow direction and position for the icecap axis (pl. 1*A*) are complemented by that determined from its north side, where the well-developed glacial-flow features in Gardners Hole fade out east of Dome Mountain and erratics of Precambrian rocks and Tertiary intrusives from Mount Holmes demonstrate northward flow from as far south as Indian Creek.

Ice flowing southwest from the icecap axis toward the West Yellowstone Basin had to flow over an escarpment 150–300 m high that extends southward from the Gallatin Range to the Madison Canyon (fig. 32). The direction of glacial scour and striations on this rim and down the tuff dipslope on its west side indicates that the ice surface sloped southeast (fig. 32; pl. 1*A*).

Bouldery, hummocky moraines along the south side of Cougar Creek and along the north side of Maple Creek represent the terminus of ice flowing over this escarpment. As shown in fig. 33, glacial streaming features on the north side of Cougar Creek indicate southwesterly flow oblique to Cougar Creek, a direction confirmed by abundant erratics of Precambrian gneiss and Tertiary intrusive rocks from the Mount Holmes area to the northeast.

Moraines north of Maple Creek are fresh, locally bouldery, and remarkably continuous (fig. 34; pl. 1*A*; Pierce, 1973a; Waldrop, 1975a); they can be walked for about 10 km to the northeast from near the Cougar Creek–Maple Creek confluence, to where a lobe of ice went down a tributary of Gneiss Creek (fig. 34). The moraines are not traceable for the next 2 km across a steep talus of Eocene volcanic rock. Farther north, in a flat-bottomed valley, several horseshoe-shaped, sharp morainal crests, with closed depressions containing

FIGURE 31.—Oblique aerial view west through Fawn Pass in the Gallatin Range. The broad U-shaped pass was shaped by ice spilling westward across the range. Pinedale terminal moraines lie 11 km west of the pass (pl. 1*A*) beyond the low meadows visible through the pass. The sedimentary bedrock at lower right was smoothed by glacial scour. Photograph by W. B. Hall.

ponds, occur just inside the Pinedale ice margin.

The contrast between terrain last occupied by Bull Lake ice and that last occupied by Pinedale ice can be observed along the Gneiss Creek trail between Madison Canyon and Gneiss Creek. The Bull Lake terrain is subdued, loess mantled, and not bouldery; grasses and sagebrush commonly dominate trees (figs. 33, 34). The Pinedale terrain, on the other hand, is primarily forested and has sandy soils, and boulders are common at the ground surface.

Soil profiles on Pinedale moraines range from ones so weak that the C-horizon occurs at a depth of 15 cm (see fig. 10, locs. 8 and 9) to those profiles in which a cambic B-horizon extends to depths of 0.4 to 0.6 m but in which little or no texture or soil structure has developed.

Melt-water channels extend from the Pinedale moraines well out onto the flats of West Yellowstone Basin. At Maple Creek they breach morainal ridges in several places; one channel about 200 m wide can be traced for 3 km from the moraines. Another outwash channel extends southward from the southernmost margin of the Cougar Creek moraines (fig. 33). The upper end of this channel (fig. 33) terminates at a steep, north-facing ice-contact escarpment 30 m high, below which are slightly younger, sharp, small to medium-sized morainal ridges. The downstream end of this channel forms an outwash fan similar to and abutting the fan emanating from the Pinedale terminal moraines of the Madison River lobe to the south (fig. 33); it shows the age equivalence of these moraines.

The absence of Pinedale terminal moraines between

FIGURE 32.—Oblique aerial photograph looking south, showing glacially scoured uplands south of Gallatin Range, indicating ice flow southwest toward the West Yellowstone Basin. Arrows indicating direction of ice flow are nearly at right angles to the strike of the sedimentary bedrock, which is from lower right to upper left.

Maple Creek and Cougar Creek is attributed to burial by Pinedale and younger alluvium along an active downwarp. At the mouth of the Cougar Creek valley, sandy alluvium is presently burying marsh vegetation, and Pinedale moraine crests become inundated by modern alluvium as they are traced down Cougar Creek. Swamps on the floor of the West Yellowstone Basin just west of Cougar Creek and in Madison Canyon suggest that the area has been downwarped below the water table. The course of Cougar Creek suggests it is flowing along an active downwarp, for it flows northward near the base of a dipslope on tuff. If downwarping were not occurring, alluviation from the streams flowing off the dipslope would be expected to build west-sloping fans, displacing the creek westward from its present course. This subsidence is probably part of the same structural process that has tilted the 600,000-year-old tuff into the West Yellowstone Basin.

At the mouth of Madison Canyon, the western boundary of Pinedale ice is delimited by terminal moraines (pl. 1*A*, fig. 33; Alden, 1953; Richmond, 1964a; Waldrop and Pierce, 1975). These moraines are less bouldery than those along Cougar and Maple Creeks, probably because they are composed almost exclusively of rhyolite. Several moraine crests are broken in places by melt-water channels, which can be traced for about 3 km onto the alluvial plain of the West Yellowstone Basin. Soils are weakly expressed in the rhyolitic parent material of these moraines (fig. 11).

OBSIDIAN-SAND OUTWASH PLAIN

A large alluvial deposit covers about 250 km^2 of the West Yellowstone Basin and is reported to average about 25 m in thickness (Richmond, 1964a, p. 227). This deposit has been variously described as a large terrace postdating the moraines described here as Bull Lake (Alden, 1953, p. 178), as having formed during the Bull Lake–Pinedale interglacial interval (Richmond, 1964a; U.S. Geol. Survey, 1972a), and as Bull Lake outwash (Witkind, 1969, p. 58). I believe these deposits to be Pinedale outwash, deposited principally by glacial-outburst floods[3] upon release of ice-dammed lakes.

Glacial-outburst floods are not only possible but likely, given the topography and the distribution of Pinedale ice and thermal areas in the western part of the park (L. J. Muffler, oral commun., 1970). Ice advancing against the bedrock buttress formed by National Park Mountain at the northeast end of the West Yellowstone rhyolite flow may have temporarily dammed lakes in the geyser basins (fig. 35). Geothermal waters in the Upper, Midway, and Lower Geyser Basins now supply enough heat to melt 4.5 m of ice a year over the entire area of the Geyser Basins (Fournier and others, 1970). Geothermal heat could have formed a lake 100 km^2 in area and 100 m deep by melting glacial ice there. Failure of the ice dam by either floating or melting would have resulted in catastrophic flooding down the Madison Canyon and out into the West Yellowstone Basin (fig. 35).

Obsidian-rich rhyolite forms nearly all the bedrock in the west-central part of the park. This rhyolite breaks down to sand and fine gravel, and was the main source for the detritus forming the obsidian-sand plain. As the floodwaters spread out into the West Yellowstone Basin, either a flood-fan or flood-delta formed at the upstream end of the basin. Without ponded waters in the West Yellowstone Basin, a flood-fan would accumulate. A possible dam for holding waters in the West Yellowstone Basin is the glacier that deposited the youthful-looking moraines across the Madison valley at Beaver Creek just 1-5 km downstream and about the same altitude as Hebgen Dam. Another possible ponding mechanism is hydraulic damming at

[3]Although this suggested origin of the obsidian-sand plain has not been published previously, a number of geologists, including W. B. Myers in his field studies in 1959, and L. J. P. Muffler, W. B. Hall, and G. D. Fraser (oral commun., 1973) have considered it. Discussions with them and with D. E. Trimble, H. E. Malde, R. L. Christiansen, and W. B. Hamilton provided incentive for presenting this interpretation. At a field conference in 1960, G. M. Richmond (written commun., 1973) discussed a glaciovolcanic-flood origin associated with emplacement of the West Yellowstone rhyolite flow.

FIGURE 33.—Pinedale terminal moraines along Cougar Creek and the Madison River. As shown in figure 10 (locs. 8 and 9), a thin soil with little or no loess is present on the Pinedale moraines, PM. Arrows, ice flow direction. Note glacial-scour features indicating Pinedale flow to the southwest, oblique to Cougar Creek. Outwash channel, OC, is bracketed by single-barb arrows and extends southwestward from Pinedale moraine. BM, large meadow area, Bull Lake ground moraine and kame gravels mantled by loess. Along Madison River, Pinedale terminal moraines are locally breached by melt-water channels that extend for at least 3 km west of the moraines. Stereo-aerial photographs by AMS, 1954.

the narrow outlet of the basin near Hebgen Dam (fig. 35). In either case, the obsidian-sand plain could then have accumulated as an extensive, low-gradient flood-delta similar to that in the Pocatello and Portland areas (Trimble and Carr, 1961; Malde, 1968; Trimble, 1963). Hydraulic damming does not appear realistic upon comparison of the probable volume of a lake in the geyser basins with the size of the West Yellowstone Basin (fig. 35).

The obsidian-sand plain grades headward into Pinedale outwash displaying channels that emanate from a Pinedale terminal moraine complex. Thus, it ties physically to Pinedale outwash. Contours show that at the mouth of Madison Canyon a large fan sloping up to 1° heads in Pinedale moraines and grades northwest into a nearly planar surfaced sand plain sloping only 0.1° (fig. 35). The deposit is dominantly obsidian clasts from rhyolite flows in the Madison

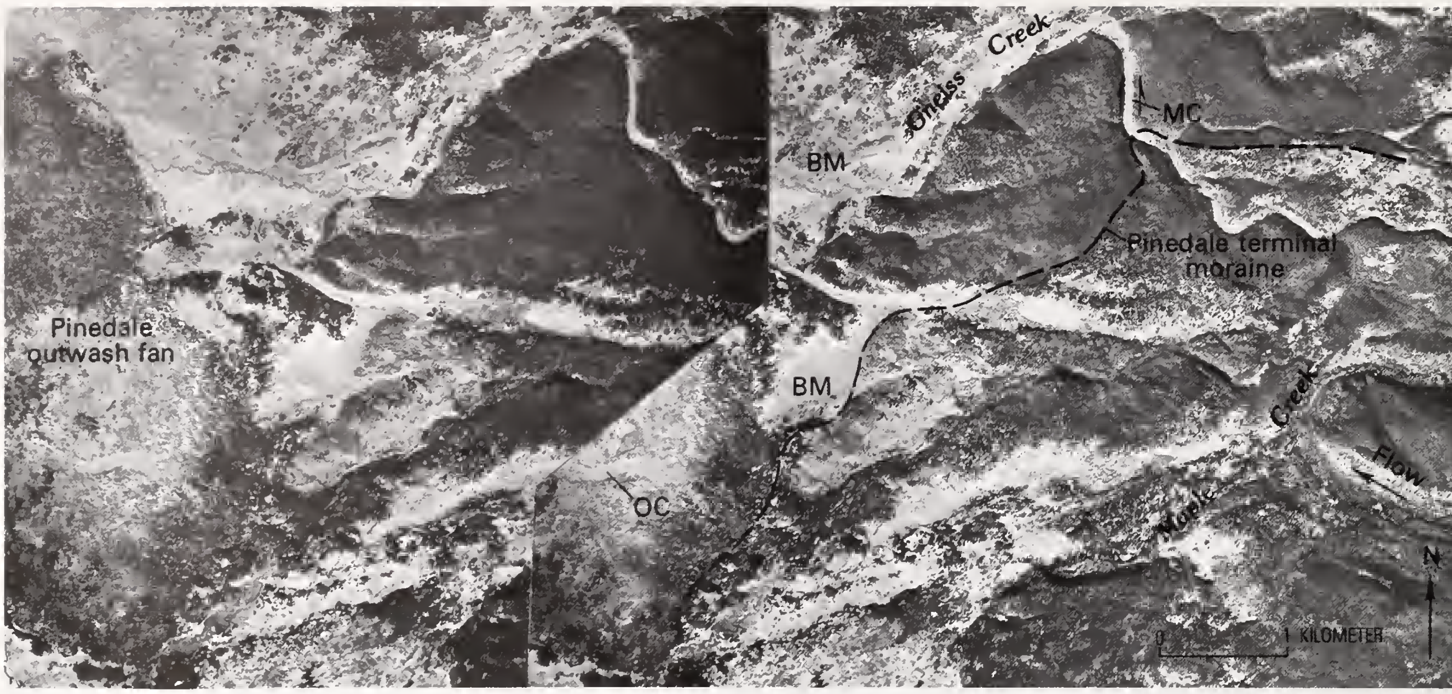


FIGURE 34.—Pinedale terminal moraine on north side of Maple Creek. Outer margin of Pinedale ice shown by dashed line. Note outwash channel, OC, from Pinedale moraine 1.6 km west of terminus. The moraine protrudes down a melt-water channel, MC, into the Gneiss Creek drainage. Stereo-aerial photographs by AMS, 1954.

River drainage, especially the West Yellowstone rhyolite flow. Grains of quartz and sanidine and fragments of stony rhyolite from both flows and tuffs accompany the obsidian clasts. The deposit coarsens eastward toward Madison Canyon. The fan gradient is so low that Alden (1953, p. 178) thought that it might have been deposited in ponded water. This low gradient apparently relates to exceptionally large discharges and velocities and perhaps ponded waters in the West Yellowstone Basin at the time of deposition.

Along Cougar Creek and to the north, the proportion of obsidian diminishes, and stony rhyolite from the tuff immediately to the east becomes dominant. Apparently, Cougar, Maple, and Gneiss Creeks aggraded in response to the rise of base level as the Madison River built the obsidian-sand plain.

Sedimentary characteristics of the obsidian-sand deposit suggest rapid deposition by sheet-flood. In several areas where bedding could be observed, the deposit consists of sheetlike beds 15–60 cm thick in which planar cross-strata are inclined at both high and low angles to the bedding (fig. 36). Most of the deposit is fine gravel to coarse sand, but silt-impregnated zones 2 cm or so thick commonly separate crossbedded layers. Away from the fan head, the deposit contains remarkably little coarse gravel, suggesting that not much coarse material was available; what little coarse gravel was present evidently was deposited within a few kilometers of the fan head.

Two other sedimentary characteristics distinguish the obsidian-sand plain from many other alluvial deposits. First, the crossbeds of fine gravel generally have openwork structure, similar to that in flood deposits of the channeled scablands of Washington (Baker, 1973, p. 36, fig. 26). The voids between the clasts are not filled with finer material and the deposit has very little cohesion. On both artificial and natural exposures, the deposit has generally slumped, and bedding characteristics are difficult to observe or photograph. Second, the well-developed cut-and-fill sedimentary structures common in most alluvial deposits are less developed, a characteristic also observed in historic flood deposits (McKee and others, 1967, p. 832).

One problem with a flood origin for the obsidian-sand plain is the lack of conspicuous large boulders protruding from the surface of the deposit. However, no exposures were observed from the fan head almost to U.S. Highway 191, where the largest boulders would be expected. "Pudding stones" and cobbles occur in the sand 15–20 km from the fan head. The bedrock source of the deposit tends to disintegrate into mostly fine gravel, sand, and silt; perhaps the scarcity of boulders is due to this characteristic of the bedrock.

Cracks in obsidian pebbles collected from 2 to 12 m below the surface of the obsidian-sand plain have

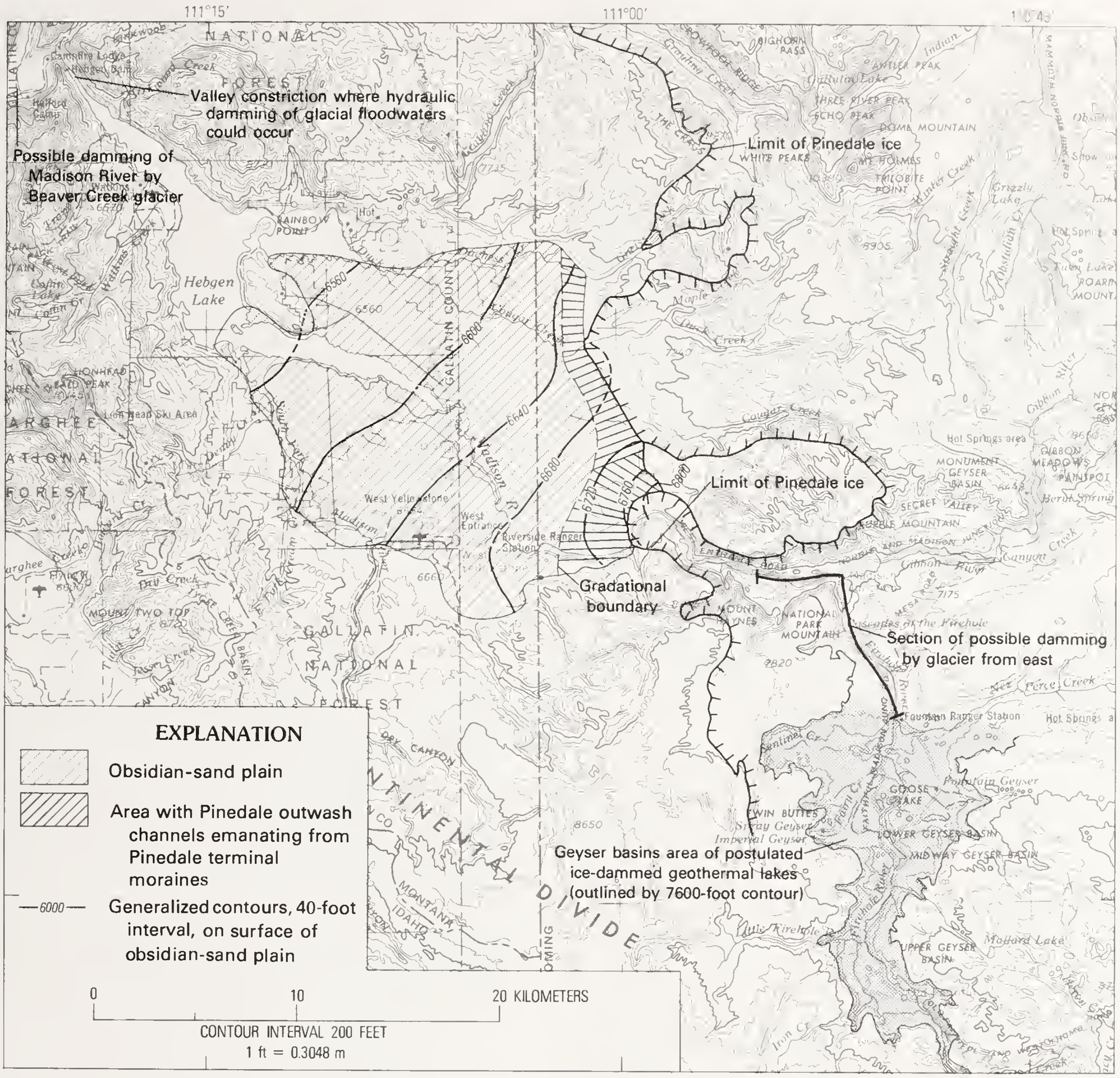


FIGURE 35.—Major elements in a glacial-flood hypothesis for the origin of the obsidian-sand plain. As shown by generalized contours, the obsidian-sand plain is a very gently sloping fan that steepens towards its source at the mouth of the Madison Canyon. Mapping of the Pinedale glacial limit from Pierce (1973a) and Waldrop and Pierce (1975). Mapping of the obsidian-sand plain is from Richmond (1964a), Pierce (1973a), Waldrop and Pierce (1975); and Waldrop (1975a). Forty-foot contours generalized from U.S. Geological Survey 1:62,500, West Yellowstone, Madison Junction, 1958. Base from AMS 1:250,000 Ashton, 1960.

hydration rinds of the same thickness as those in the Pinedale terminal moraines. The character of the cracking is also the same as that in the terminal moraines.

Soil profiles developed on the obsidian-sand plain are weak and are similar to those on other sandy deposits of Pinedale age. More silt occurs in the upper 50 cm than at greater depths, in part probably because windblown silt has been washed down into the deposit.

EAST-WEST ICECAP AXIS ABOVE THE OBSIDIAN CLIFF FLOW

Between the Gallatin and Washburn Ranges the axis of the plateau icecap trended east-west, crossing the southern third of the Obsidian Cliff rhyolite flow near Lake of the Woods (pl. 1*A*). The evidence for locating this ice divide seems equivocal at first, for as one approaches the ice divide or "ice source," the evidence for glaciation, such as scour features, erratics, and glacial

FIGURE 36.—Obsidian-sand-plain sediments. Horizontal beds about 15 cm thick have both subhorizontal and steeply inclined crossbedding. The inclined crossbeds of pea-sized gravel are generally openwork (have no matrix). Scale is 18 cm long. Photograph by W. B. Hamilton, 1959.

deposits, becomes weaker and weaker, and the terrain has a less freshly glaciated appearance. But this less glaciated aspect of the terrain may be explained if one considers the conditions prevailing beneath an icecap axis (fig. 27). Little or no flow occurs immediately beneath an icecap axis where terrain is nearly horizontal, so scour features would not be formed. Erratics would be absent, for the outward-flowing icemass would prevent their introduction. Erratics are extremely rare east from the Obsidian Cliff flow to the brink of the Lava Creek Canyon. They are more than two orders of magnitude less common than in moraines from ice sources in the Gallatin Range, Washburn Range, or Beartooth uplift. Glacial scour features on the Obsidian Cliff flow show an interesting progression. On the southern part of the flow, original flow-top depressions form small marshes in the lodgepole pine forest. Northward the marshes disappear and faint evidence of scour becomes apparent. Further north, on the Crystal Springs flow (pl. 1*A*), scour features are better expressed.

Southward from the location of the icecap axis, glacial scouring, streaming, and molding features become apparent just north of Norris Geyser Basin, where glacial flow features are alined toward the southwest (pl. 1*A*). Flow from this region was directed toward the north rim of Madison Canyon.

Near its axis the icecap was as much as 800 m thick, as determined by projecting the ice surface from either known or minimum heights of ice in the Gallatin Range to the west, in the Washburn Range to the east, and on nunataks to the north and south.

The history of fracturing of the Obsidian Cliff flow furnishes some unique and independent information regarding the Pinedale icecap (Friedman and others, 1973). Hydration rinds on fractures appear to cluster into three groups, averaging 16.2, 14.5, and 7.9 mm in thickness. The oldest generation of cracks (thickest hydration) was formed during cooling of the flow, which is dated by K-Ar as 183,500±3,000 years old (J. D. Obradovich, written commun., 1979, 1977 K-Ar constants). The thickness of the rinds on the other two sets of cracks corresponds closely with that on glacially induced cracks in obsidian pebbles in Bull Lake and Pinedale end moraines of the West Yellowstone area. These two sets of fractures in the Obsidian Cliff flow are postulated to be due to glacial loading by Bull Lake and Pinedale icecaps. The Pinedale icecap is thought to have been thicker than that of the Bull Lake in the northern Yellowstone area (fig. 17), so that its load would have been greater than that of the Bull Lake, and would have caused new fracturing of the flow.

ICE DIVIDE SOUTH OF THE WASHBURN RANGE

During full-glacial conditions an ice divide existed south of the Washburn Range. The nearly due north direction of flow across the range indicates that the icecap had an east-west-trending axis. This axis spanned the area from Wolf and Grebe Lakes to just north of Canyon Village. The northeasterly trending topographic grain apparent on aerial photographs northeast of Canyon Village is probably due to glacial flow around the east end of the Washburn Range (fig. 27).

From Canyon Village, the icecap axis trended easterly to the Mirror Plateau. As shown on plate 1*A* above the Mirror Plateau, the east-west axis of this icecap abutted both the north-south axis of the icecap over Yellowstone Lake and the southwesterly axis of the icecap in the upper Lamar drainage.

ICE DIVIDE IN THE UPPER LAMAR DRAINAGE

The exact location of the upper Lamar ice divide is uncertain (pl. 1*A*) because of the absence of glacial-scour features except along the drainage divide itself. Glacial-scour features along the east boundary of the park at the headwaters of the Lamar River indicate that ice flowed eastward across the present drainage divide into the drainages of the Clarks Fork and Shoshone Rivers.

Large-scale scour features along the drainage divide are best recognized on wide flats such as the grassy uplands southeast of Hoodoo Peak, at Bootjack Gap, and at Canoe Lake (an ice-scoured rock basin on the divide (pl. 1*A*; Pierce, 1974b)). All the striations noted were on the Eocene volcanic bedrock. The amount of

postglacial weathering, frost activity, and the freshness of glacial scour are all similar to Pinedale features observed elsewhere.

ICE DIVIDE ON THE BEARTOOTH UPLIFT

The ice divide of the large icecap on the Beartooth uplift was mostly south of the present drainage divide (pl. 1*A*). Aerial photographs show all the passes along this drainage divide to be intensely scoured by ice. Ground examination of the divide revealed glacial striations and scour features showing northward flow. Geographic relations also suggest that the ice divide was south of the topographic divide. First, the distance to low elevations is less north of the divide than to the south, thus providing both a steeper gradient and faster melting for glaciers flowing north from the divide. Second, the congestion caused by the nearly centripetal flow of ice from the different sources of the northern Yellowstone glacier (pl. 1*A;* fig. 5) would lead to higher ice levels south of the divide on the Beartooth uplift.

ICE DIVIDE BETWEEN SODA BUTTE AND CLARKS FORK GLACIERS

The mountain mass culminating in Pilot and Index Peaks constituted a "prow" between the Clarks Fork and Soda Butte glaciers. The location of the ice divide there was estimated by plotting the flow directions of about 50 glacial-scour features, and then drawing a line between diverging directions of flow separating ice flow down Soda Butte Creek from that down Clarks Fork (pl. 1*A*).

The source area for the ice flowing down Soda Butte Creek and Clarks Fork was a large spoon-shaped upland above 3,050 m (10,000 ft) in altitude rising to Granite Peak, the highest point in Montana, at an altitude of 3,901 m (12,799 ft). The glacier that went down the Clarks Fork terminated at the mouth of the Clarks Fork Canyon about 70 km from its icecap source (W. G. Pierce, 1965a, b).

CHANGING GLACIAL DYNAMICS DURING WANING GLACIAL CONDITIONS

Because thick icemasses from different source areas abutted against each other during full-glacial conditions (pl. 1*A*), deglaciation of the northern Yellowstone area did not consist simply of glacial recession, interrupted by temporary halts and moraine building. Relative differences in source-area productivity under waning glacial conditions resulted in flow of ice from some sources into the full-glacial domain of other ice sources. As the ice was generally on relatively open terrain rather than confined by valley walls, flow could occur in a variety of directions. This section first describes glaciers that receded in a relatively uncomplicated manner and then discusses the complex interrelations of icemasses which formed the northern Yellowstone outlet glacier.

SIMPLE UPVALLEY RECESSION

GALLATIN VALLEY GLACIER

The Gallatin Valley glacier is a good example of relatively simple "normal" recession. The unusually well developed moraine sequence on the valley floor of the Gallatin River shows at least five standstills or readvances during a general recession from the Pinedale maximum prior to late Pinedale time, each with associated outwash deposits (fig. 37).

All moraines in this sequence are assigned to the Pinedale on the basis of weathering rinds, preservation of morainal morphology, degree of soil development, and degree of breaching by the Gallatin River. Also, a layer of volcanic ash 0.5 cm thick lies 0.5 m above the base of a core 5 m long, which was taken from sediments in the kettle pond inside the outermost pair of end moraines. (See figure 37.) R. E. Wilcox (written commun., 1971) reported that the ash has the petrographic properties of either the Glacier Peak ash, dated as about 12,000–13,000 years old (Fryxell, 1965; Mudge, 1967), or ashes from Mount St. Helens, dated as between 8,000 and 12,000 years (Mullineaux and others, 1972). Pollen from the lower 1.5 m of core indicates a cold climate similar to that during deglaciation elsewhere in the Yellowstone area in the interval 13,500–11,500 years B.P. (R. G. Baker, oral commun., 1972). The ash occurs in the lower part of this pollen interval. The association of the ash with a typical deglacial pollen sequence indicates an age of >11,500 years for the ash. Thus the ash is either the Glacier Peak ash or the older one of the suite of Mount St. Helens ashes.

The fact that this kettle was undrained and only partly filled with sediment indicates that it is no older than the last major glaciation—the Pinedale. Although an area of about 2 km^2 drains into this kettle, which is 150 m in diameter, it is less than half filled.

Weathering rinds on andesitic intrusive rocks derived from the Gallatin Range were measured to ascertain differences from the outer to the inner moraines. The rinds are better developed beneath places where lichens are growing or where the effects of glacial abrasion are not apparent, so the following procedure was used in measuring rinds: stones of andesitic intrusive rock well imbedded in soil along a moraine crest were

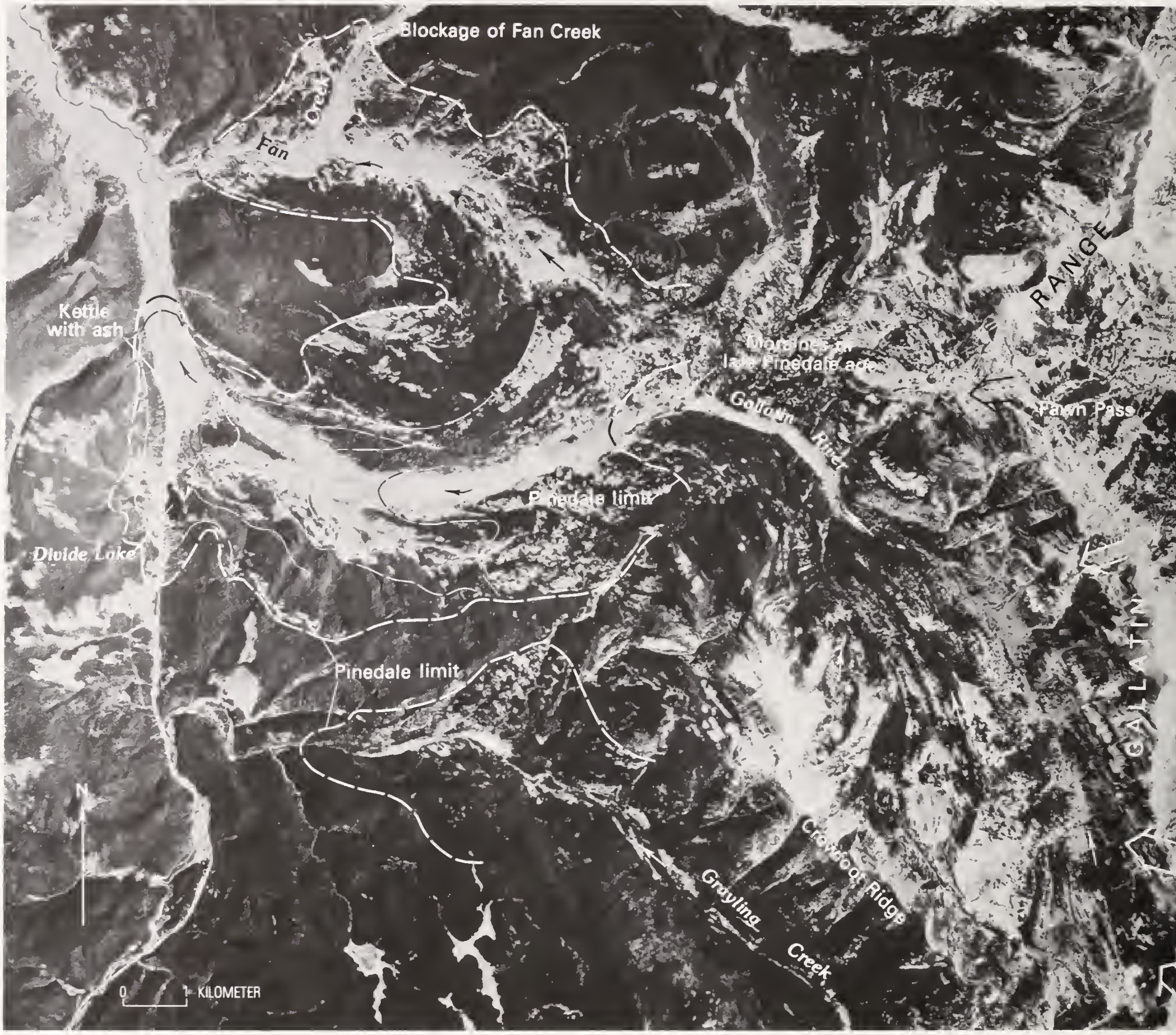


FIGURE 37.—Aerial photograph showing glacial features on the west side of the Gallatin Range. Ice flowing through passes along the Gallatin Range divide (open arrows) combined with glaciers from local cirques (at lower right) to form a glacier in the Gallatin valley large enough to spill northward and block Fan Creek. The main stem of the glacier moved down the Gallatin valley to the trench followed by U.S. Highway 191, whereupon one lobe flowed downstream (north) and another moved south into the drainage of Grayling Creek. Along the Gallatin River valley well-defined moraines delimited by dashed lines show six subdivisions of Pinedale moraines downvalley from moraines of late Pinedale age. All the moraines have about the same morphologic expression, degree of soil development, thickness of weathering rinds, and degree of breaching by streams. Photograph by NASA, 1969.

pried out and broken, then rinds were measured on the underside of the stones inward from smooth, convex surfaces having a glacially abraded appearance. From the outermost to the innermost major moraine crest shown in figure 37, most of the rinds measured were thinner than 0.5 mm—the minimum thickness I could clearly measure on rock of this texture.

Sites can be found on each of the moraines, from the innermost to the outermost in both the Gallatin River and Fan Creek drainages, where the strongest morphological expression is approximately the same in terms of boulderiness, steepness, and hummockiness. However, much of the moraine morphology throughout the sequence is subdued, apparently

because the till contains a large amount of fine-grained material from shaly bedrock and only a few large boulders.

Soil profiles were examined in 15 soil pits 1.2–1.8 m deep on the moraine crests and outwash of the Gallatin valley Pinedale moraines, from the outermost to the late Pinedale. The tills in these moraines are about one-fourth clay and are calcareous. In an initial search for Bull Lake moraines in the area, differences in soil development were thought to represent the Pinedale–Bull Lake age difference. Weakly to moderately developed soil structure and leached B-horizons about 0.6 m thick were found in the soils in the outer three moraines, while soils in moraines farther upvalley appeared to be less developed. However, more detailed investigation revealed soils comparable to those on the outer moraines on all of the upvalley moraines, including those of late Pinedale age. Thus, the soils data are consistent with the morphology of the moraines and indicate that the moraines are all of Pinedale age. The stronger than normal development of post-Pinedale soil profiles in the Gallatin valley seems to be related to moderate clay and carbonate content, combined with saturation by heavy snowmelt and a lush grassland vegetation. Soils on similar, weakly consolidated Pinedale deposits east of the Gallatin Range also have similar profile characteristics (Pierce, 1973a).

GLACIERS IN THE VALLEYS OF COUGAR, MAPLE, AND GRAYLING CREEKS

Along Cougar and Maple Creeks no significant morainal deposits occur from the end moraines for about 15 km up the tuff dipslope to the top of the dipslope. The top of this dipslope formed a threshold across which ice only about 300 m thick flowed under full-glacial conditions. When the ice level was lowered by only about 200 m (one-third of its full-glacial thickness on the plateau), flow across the threshold effectively ceased. Thus ice west of this threshold retreated more rapidly than normal.

A similar threshold existed at the head of the Grayling Creek glacier. Most of the ice forming the 11-km-long glacier in the valley of Grayling Creek entered the drainage around the southwestern flank of the Gallatin Range and through several gaps along the Gallatin Range crest (fig. 1). Several end moraines were formed in the valley of Grayling Creek during full-glacial conditions when ice was as much as 450 m thick. But extensive kame-gravel deposits, which extend for a distance of about 10 km upvalley, suggest that after the end moraines were deposited the glacier abruptly stagnated. This stagnation occurred because the glacier was dependent upon inflow from the nearby icecap, which effectively ceased when the ice surface was lowered by about 200 m.

PHASE RELATIONS BETWEEN AREAS FEEDING THE NORTHERN YELLOWSTONE OUTLET GLACIER IN PRE-DECKARD FLATS TIME

The northern Yellowstone outlet glacier flowed nearly undiminished down the Yellowstone valley while glaciers in tributary valleys were significantly diminished. A similar relation on a larger scale occurred in Puget Sound lowland of Washington State during the last major glaciation, where by the time a glacial lobe from a large Canadian icecap reached its maximum, local alpine glaciers in the Cascades had already greatly decreased in size or disappeared (Crandell, 1965, p. 346). Something similar occurred in the southern Rocky Mountain trench of British Columbia, where in the later part of the last glaciation, trunk-glacier ice advanced into tributary valleys that had previously supplied ice to the trunk glacier (Clague, 1975).

BACKFILL IN TOM MINER BASIN

By the time the outlet glacier pouring through Yankee Jim Canyon had thinned only 20 percent from its full-glacial thickness (900 m to 720 m), local glaciers in Tom Miner Basin had already largely disappeared. The outlet glacier backfilled into Tom Miner Basin and deposited multiple moraines that slope into this basin (fig. 38). Kame gravels and ice-dammed lake sediments were deposited in the central part of the basin where the local drainage was blocked by the outlet glacier. The continuity of moraines and abundant erratics of Precambrian rocks show that the outlet glacier extended into the basin for a distance of 9 km from the Yellowstone River. Precambrian bedrock does not occur in Tom Miner Basin; hence, these erratics were mostly derived from the Precambrian outcrops in Yankee Jim Canyon.

Although the backfill in the Tom Miner Basin generally has moderate to subdued morphology characteristic of the wet environment of backfills, the morainal morphology is quite strong in places, with steep slopes and undrained conical depressions.

A reconstruction of glacier gradients at the time of maximum backfilling of Tom Miner Basin indicates that the northern Yellowstone outlet glacier terminated down the Yellowstone valley at or near the Chico moraines.

FIGURE 38.—Oblique view to south of morainal backfill in Tom Miner Basin. Moraines sloping from left to right into Tom Miner Basin were deposited by the northern Yellowstone outlet glacier where it debouched from Yankee Jim Canyon (left of photograph). The outlet glacier carried erratics of Precambrian rocks far into Tom Miner Basin. When the highest level of the backfill was deposited, glaciers in Horse Creek valley, which lies beyond the right side of the photograph, had retreated well upvalley. Photograph by W. B. Hall.

Horse Creek drains northward into the east end of Tom Miner Basin. Glaciers flowing down Horse Creek received ice both from five cirques at the head of Horse Creek and from a Pinedale icecap on the upland (pl. 1*A*) at the head of this drainage. The distribution of erratics and moraines indicates that the outlet glacier pushed across and up Horse Creek to a position only 5 km downvalley from the cirques. No end moraines from Horse Creek occur on the backfill deposits, and no end moraines were noted along Horse Creek to within about 3 km of the cirques. Because no terminal moraines from the Horse Creek glacier are present on top of the outlet glacier moraines, the backfill is younger than the Pinedale maximum of Horse Creek. At full-glacial conditions, the Horse Creek glacier joined the outlet glacier and was probably more than 300 m thick (pl. 1*A*). Subsequently, when Horse Creek was largely deglaciated, the outlet glacier was still nearly at its full-glacial size. I estimate that when the northern Yellowstone outlet glacier was still at more than 90 percent of its maximum volume, Horse Creek glacier was at less than 50 percent of its maximum volume.

Soils also indicate a Pinedale age for the till, kame gravels, and lake sediments backfilling Tom Miner Basin. Most have weakly developed profiles, generally only a cambic B-horizon 0.2–0.5 m thick over a weakly to moderately developed Cca horizon 0.2–0.4 m thick. Soil structure and clay enrichment of the B-horizon generally are absent to weakly developed.

The Pinedale backfill in Tom Miner Basin is important in evaluating criteria for the size of the Pinedale outlet glacier near Gardiner. In 1968 I thought that the Deckard Flats ice margin (a Pinedale recessional phase) (pl. 2) represented the limit of Pinedale ice. This interpretation was based mainly on well-developed soil profiles, two of which are located and briefly described in Pierce (1973b, profiles A and B) that seemed to require a Bull Lake age for glacial deposits located above the Deckard Flats limit north of Gardiner, Mont. But in 1970 I found that the Pinedale moraines that backfilled Tom Miner Basin reached an altitude of more than 2,200 m (7,200 ft) on the downvalley side of Yankee Jim Canyon, which is higher than the anomalous soil profiles near Gardiner. Thus, the well-developed soil profiles occur in an area glaciated by Pinedale ice, and do not demonstrate a pre-Pinedale age of the underlying deposits. Apparently soil development was enhanced by swelling clays in these deposits.

OTHER TRIBUTARY GLACIERS NORTH OF YELLOWSTONE NATIONAL PARK

The lower parts of both Emigrant and Sixmile Creeks were free of ice when the northern Yellowstone outlet glacier had receded no more than 5 km from its terminus. Unconsolidated, horizontally bedded lake sediments in the lower parts of these valleys show that both valleys contained lakes blocked by the outlet, or piedmont glacier (Montagne, 1970, p. 15; pl. 2). The valley shapes of Emigrant and Sixmile Creeks also suggest that glaciers from these valleys contributed little to the trunk glacier. As noted by John Montagne (1970, p. 15),

> Strangely, Sixmile and Emigrant Creeks emanate from the Beartooth front in deep "V" shaped canyons, showing little glacial modification * * *. Because these valleys lie beneath towering peaks with apparent ice gathering potential, it is possible that piedmont ice floated the valley ice here thus preventing erosion. Major valleys beyond the piedmont terminus show normal glacial profiles.

Cinnabar Creek and Mol Heron Creek, south of Tom Miner Basin (pl. 1*A*), drain a section of the glaciated crest of the Gallatin Range about 13 km long. Glacial flow from the valleys of these creeks was dominated by the outlet glacier throughout recession. Weed (1893, p. 18, pl. 1) observed erratics high on the slopes of the southern portal of Yankee Jim Canyon, "whose only source is Cinnabar Mountain." Subsequent field studies in the area have confirmed the position and probable source of these erratics (John Montagne, oral commun., 1972). As shown on plate 2, the flow trajectory of the outlet glacier represents a boundary across

which local glaciers did not flow when the erratics were emplaced, thus indicating the dominance of the outlet glacier over the local glaciers. The erratics occur about 180 m below the upper limit of Pinedale ice, and were probably emplaced slightly after full-glacial time. The flow trajectory shown by the erratics is consistent with glacial scour features that parallel the Yellowstone valley. Recessional lateral moraines of the outlet glacier extend back into the drainage of Cinnabar Creek at altitudes between 2,200 and 1,890 m (7,200 and 6,200 ft).

In the Bear Creek area, large erratics of Precambrian granitic rocks occur north of the Yellowstone River, as high as 900 m above the river, at localities such as Palmer Mountain, Parker Point, and the southern spur of Sheep Mountain (pl. 2). These erratics were not deposited by ice from Bear Creek, for no granitic source is known there (Seager, 1944, p. 37; Wedow and others, 1975; G. D. Fraser, H. J. Prostka, and D. L. Gaskill, oral commun., 1972). Their likely source is the Black Canyon of the Yellowstone or areas farther upstream. This flow trajectory establishes a boundary across which Bear Creek ice did not flow at the time the erratics were emplaced (pl. 2). Thus when the Pinedale outlet glacier was at least 70 percent of its maximum thickness, it flowed across the valley of Bear Creek. At this time, any outflow from the Bear Creek glacier could only have been high on the valley wall above the position of the granitic erratics.

ICE IN CENTRAL NORTHERN YELLOWSTONE NATIONAL PARK

As the outlet glacier terminus receded towards Gardiner, changes in flow pattern accompanied lowering of ice heights. Once the plateau icecap had diminished to the extent that it could not flow over the Washburn Range (pl. 1*A*), an incursion of ice from the Beartooth uplift pushed around the northwest flank of the range and flowed to the southwest, leaving erratics of Precambrian rocks and striations below an altitude of 2,700 m (8,800 ft) (pl. 2). Striations on the ridges northwest of Cook Peak show that the flow direction during this incursion of Beartooth ice was 130° different from full-glacial flow. This near reversal in flow probably occurred because the major icecap on the Beartooth uplift could sustain glaciers under a waning glacial climate better than the icecap on the plateau 600 m lower. Nearly stagnant ice seems to have remained on the plateau between the Gallatin and Washburn Ranges, inasmuch as erratics of Precambrian rocks are absent in the heart of that area, although they do occur near its margin (pls. 1*A*, 2).

At altitudes of about 2,300 m (7,500 ft), ice-dammed lake sediments and kame gravels along Lava Creek (Pierce, 1973b) and erratics of Precambrian rocks just west of Lava Creek (pl. 2) indicate continued incursion of ice from the Beartooth uplift into the full-glacial domain of the plateau icecap.

The domain previously occupied by the plateau icecap between the Gallatin and Washburn Ranges was also invaded by ice from the Gallatin Range. As ice levels receded below Bunsen Peak, the flow changed from northerly to northeasterly. Erratics below an altitude of 2,450 m (8,000 ft) on Bunsen Peak indicate that ice from the Gallatin Range was able to advance eastward against a much-diminished icecap between the Gallatin and Washburn Ranges.

DECKARD FLATS READJUSTMENT

The Deckard Flats readjustment is defined by moraines and other glacial features that formed when the northern Yellowstone icemass had diminished to between 30 and 50 percent of its full-glacial volume. The term readjustment is used because an important aspect of this phase is that the icecaps on the Yellowstone plateau had receded to the point where they ceased to have an effect on the northern Yellowstone glacier. Thus the Deckard Flats readjustment apparently is different from a normal glacial readvance or standstill. The Deckard Flats ice margin (pl. 2) lies well beyond the moraines of late Pinedale age (pl. 3) and was mapped in detail, for it was once thought to be the limit of Pinedale ice. The boundary was mapped from Deckard Flats (pl. 2) nearly continuously for 40 km southeast to Specimen Ridge and for 30 km south to the Twin Lakes area. The position of this boundary is less certain around the south side of the Washburn Range and around the south end of the Gallatin Range.

YELLOWSTONE VALLEY

The type moraines of the Deckard Flats readjustment lie against the valley slopes at the margin of Deckard Flats, a basalt-floored bench about 350 m above the Yellowstone River 5 km east of Gardiner (pl. 2; fig. 39). A set of continuous morainal ridges extends for about 3 km downvalley to Bear Creek where meltwater channels lie outside each of the moraines. Deckard Flats ice dammed Bear Creek, resulting in the deposition of 60 m of lake silts and overlying gravels (Pierce, 1973b; Fraser and others, 1969, p. 59). At this time, the Bear Creek glacier had receded upvalley past Jardine.

At Deckard Flats, and also on the travertine bench just to the north (pl. 2), Deckard Flats moraines are commonly about 10 m high and are studded with er-

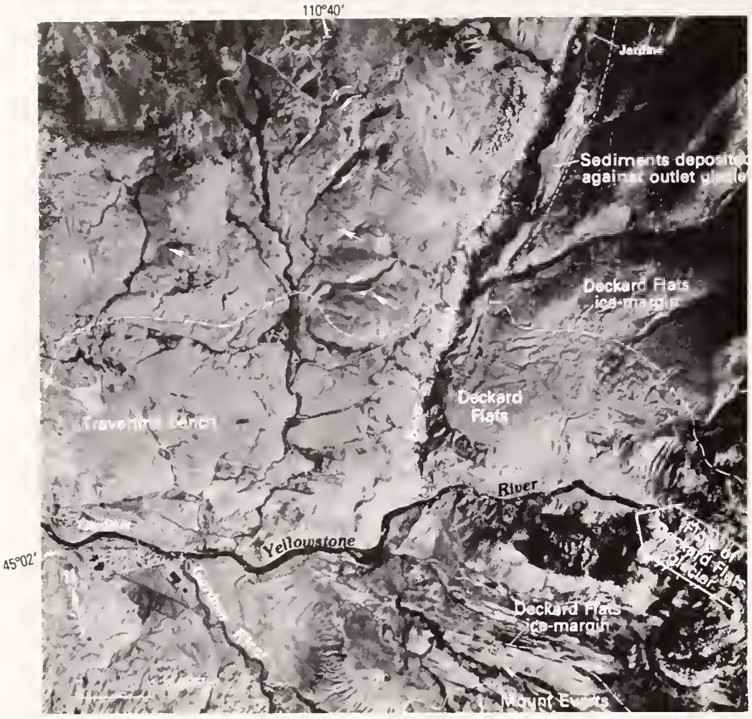


FIGURE 39.—Type area of the Deckard Flats readjustment. Several morainal ridges can be seen along the northeast margin of the bench known as Deckard Flats. At Deckard Flats time the northern Yellowstone glacier was blocking Bear Creek resulting in deposition of silt, sand, and gravel within the area bordered by dotted lines. The receding northern Yellowstone glacier that dammed Bear Creek was 350 m thick, whereas the local Bear Creek glacier had receded upvalley past Jardine, within 7 km of the Bear Creek cirques. Ice flow during full-glacial conditions (arrows) was parallel to the Yellowstone valley. Half-arrows indicate meltwater channels occupied during Pinedale recession. Aerial photograph by NASA, 1969.

ratics of Precambrian rocks, many more than 2 m in diameter. The nearest and most logical source for these granitic erratics is the Black Canyon of the Yellowstone (pl. 2).

Across the Yellowstone valley from the type locality, the Deckard Flats boundary is defined by thick morainal deposits at the base of the northern spur of Sepulcher Mountain (Pierce, 1973a, b). These deposits contain granitic erratics probably derived from the Black Canyon of the Yellowstone and carried westward across the Gardner River valley. Consequently, it appears that glaciers from the Gardner River valley (pl. 2) did not contribute much ice to the Deckard Flats terminus.

Deckard Flats end moraines on the Yellowstone valley floor apparently were obliterated by floods as much as 60 m deep that rushed down the valley in late Pinedale time. (See section entitled "Late Pinedale and younger floods.") However, the position of the lateral moraines indicates that the Deckard Flats terminus was near Reese Creek. The thick flood deposits on the south side of the river just upstream from Reese Creek may be Deckard Flats end moraines partly reworked by floodwaters.

From Deckard Flats, large lateral moraines can be traced up the north side of the Black Canyon of the Yellowstone River to the Crevice Creek area (pl. 2). Just east of Crevice Creek glacially polished and striated schist 150 m above the Deckard Flats limit is identical in appearance to that below the limit, suggesting that the duration of Deckard Flats time was short. On Mount Everts (a nunatak surrounded by Deckard Flats ice), the prominent scour features formed under full-glacial conditions trend northwesterly across the entire area of the mountain and are little modified where they cross the Deckard Flats limit, also suggesting that Deckard Flats time was short.

GLACIERS FROM THE GALLATIN RANGE

Deckard Flats ice from the Gallatin Range poured northward into the lowlands near Mammoth (pl. 2). This ice apparently joined but contributed little to the glacier in the Yellowstone valley, for, as described earlier, erratics probably derived from the Black Canyon of the Yellowstone River were carried across the mouth of Gardner River valley.

The extensive kame deposits around the hot springs at Mammoth formed just inside the Deckard Flats ice margin (pl. 2). Capitol Hill, just south of Mammoth, and other conical hills to the southeast and northeast are probably "thermal kames" formed where glacial sediments accumulated in holes and embayments melted in the ice by hot springs.

Well-developed end moraines just south of Mammoth (pl. 2) (Alden, 1928, p. 67) record a younger stand of Deckard Flats ice as it receded up the Gardner River. These end moraines, here named the Gardner River Bridge moraines, are characterized by hummocky terrain with steep slopes, morainal ridges, and numerous undrained depressions. They superficially resemble the landslides that are magnificently developed on the western valley-slopes of Mammoth hole from Bunsen Peak to north of Gardiner. Some landslide movement also has occurred in the end-moraine area, but the bulk of the area has not slid, probably because the general slope is low in this area.

At the time the Gardner River Bridge moraines were built, nearly all the ice from the Gallatin Range flowed down Sheepeater Canyon of the Gardner River and only a little came through the narrow notch at Golden Gate. Ice from the Beartooth uplift spilled down Lava Creek leaving a sharp-crested lateral moraine just below the road about halfway between Sheepeater

Canyon bridge and Undine Falls (Pierce, 1973b). Late in the construction of the Gardner River Bridge moraines, Gallatin Range ice pushed up the lower part of Lava Creek valley leaving morainal ridges that are convex upvalley, showing that ice from the Beartooth uplift had already receded from the mouth of Lava Creek.

At the Deckard Flats maximum, ice from the Gallatin Range formed a piedmont glacier on the plateau to the east. This glacier carried erratics eastward to just beyond Obsidian Creek (pl. 2). As indicated by the transport direction of these erratics and by minor scour features, a 60°–80° change of flow direction occurred in this area between full-glacial and Deckard Flats time. The eastward flow of the Gallatin Range piedmont glacier could not have been either strong or long lived, for its scour features are poorly developed and the strong northerly drumloid terrain formed under full-glacial conditions was scarcely modified even at valley mouths (fig. 22; frontispiece). About 2.4 km south of Swan Lake, crossed striations indicate northeasterly flow followed by easterly flow (Pierce, 1973b; pl. 2). Along the valley of Winter Creek, easterly glacial scour and streaming features transect the stronger topographic features formed by the previous northerly flow at nearly right angles.

The eastward limit of abundant erratics from the Gallatin Range defines the eastward limit of the Deckard Flats ice. The abundance of these erratics decreases by at least one order of magnitude within several hundred meters outside the Deckard Flats boundary, and by at least two orders of magnitude within a kilometer. The distribution of erratics in this area illustrates the point that a few scattered erratics do not necessarily indicate a single-cycle flow pattern for transport of erratics. For this area, a single-cycle flow pattern is demonstrated only by abundant erratics that constitute a significant volume of the till and that occur continuously back to the source. Ice-flow patterns similar to that of Deckard Flats time probably occurred during earlier times of glacial buildup and recession. Erratics transported eastward at such times were subsequently reworked and redistributed in full-glacial times, resulting in the sparse scattering of erratics outside the Deckard Flats limit.

Southeast from the Gallatin Range, thick till embankments and moraines lying against a gentle, northeasterly facing escarpment represent the Deckard Flats margin (pl. 2). Abundant erratics from the Gallatin Range were carried southeast by Deckard Flats ice as far as Twin Lakes. Gallatin Range erratics are also quite common southwest of the Deckard Flats limit, but till thickness and the abundance of erratics are clearly greater inside the Deckard Flats limit. The Gallatin Range erratics outside the Deckard Flats boundary were probably emplaced as follows: during either glacial buildup or deglaciation prior to Pinedale full-glacial time, erratics were transported southeastward, similar to Deckard Flats transport, and were then reworked towards the southwest during Pinedale full-glacial time.

When ice levels east of the Gallatin Range receded 50–100 m below the Deckard Flats maximum, moraines were draped around emerging nunataks between Indian and Winter Creeks (pl. 2), and around a 150-m-high ridge of Eocene volcanic rocks on the northeast side of Gardners Hole.

GLACIERS FROM THE BEARTOOTH UPLIFT

Moraine embankments, erratics, and glacial scour show that ice from the Beartooth uplift flowed southward during Deckard Flats time into areas occupied by northward-flowing ice in full-glacial time (pls. 1*A*, 2). This incursion can be explained by relatively rapid deglaciation of the Yellowstone plateau and relatively slower deglaciation of the Beartooth uplift, and requires no necessary buildup of glaciers in the Beartooth uplift.

Ice from the Beartooth uplift poured westward around the north end of Prospect Peak and deposited a nested set of moraines composed of till and kame gravels over an area of about 40 km^2 of the Blacktail Deer Plateau. Melt-water channels there indicate as many as 10 successive ice-margin positions during Deckard Flats time.

This Beartooth ice also flowed southward into the drainage of Tower Creek and across the flats of Antelope Creek, carrying erratics of Precambrian crystalline rocks and Paleozoic limestones up to altitudes of 2,620 m (8,600 ft) on the Washburn Range (pl. 2). It also left poorly developed southeasterly trending striations at an altitude of nearly 2,620 m on the north spur of Mount Washburn. The flow was at right angles to that during full-glacial conditions.

Specimen Ridge is favorably positioned to record the changes in flow between full-glacial and Deckard Flats time. By Deckard Flats time, the icecap on the plateau was considerably more diminished than the Beartooth uplift icecap, resulting in a southward incursion of ice from the Beartooth uplift. A younger set of glacial striations on Specimen Ridge transect the northwesterly full-glacial striations and indicate flow to the south at altitudes as high as 2,750 m (9,000 ft). Erratics of Precambrian crystalline rocks and of Paleozoic limestone also indicate southerly flow of ice up to altitudes of 2,750 m.

Remnants of the Yellowstone plateau icecap evidently prevented flow of Beartooth ice into the area between Mount Washburn and Amethyst Mountain. Erratics of Precambrian rocks are common between Specimen Ridge and the northern rim of the Deep Creek canyon, but none could be found in several days of traverses from Deep Creek south almost to Canyon Village (pl. 2). Erratics are common farther south in the Canyon Village area, but they were probably carried south of the Washburn Range before Pinedale full-glacial time. (See end of section on Bull Lake Glaciation.)

Ice from the Beartooth uplift seems to have poured southward across parts of Mirror Plateau during Deckard Flats time. As shown by the slope of moraine crests there, the ice surface sloped southward at an altitude of about 2,800 m (9,200 ft). At this time Amethyst Mountain was probably a nunatak, with ice from the north pouring around it into an area of nearly stagnant ice south of Specimen Ridge.

PLATEAU ICECAP SOUTH OF THE WASHBURN RANGE

During Deckard Flats time, a glacier stood against the southern base of the Washburn Range (pl. 2). The margin of this glacier can be traced to the southeastern end of the Washburn Range where it is at the same altitude as the Deckard Flats limit of Beartooth ice. It can also be traced westward to the Norris area where it is at the same altitude as the Deckard Flats limit of Gallatin Range ice. Along the southern margin of the Washburn Range this ice margin is generally represented by inwash alluvium and mass-movement deposits, which were piled against the inferred Deckard Flats ice front. Just north of Grebe and Wolf Lakes, thick kame sands composed of Quaternary rhyolitic tuff derived from the terrain to the north (pl. 1*A*) were deposited against the front of ice that occupied the area now containing the lakes (pl. 2). Farther east along the lower slopes of the Washburn Range, escarpments 5–15 m high composed of clayey volcanic debris from the Washburn Range (Pierce, 1974a) can be traced rather continuously. Here the sediment is commonly neither sorted nor stratified; it was deposited by mudflow and debris avalanches from steep slopes underlain by poorly indurated volcanic debris from the Washburn Range. Red clays occur in the bedrock and are incorporated into the poorly sorted deposits along the base of the range. These clays apparently account for the moderately good soil "development" in this area. Clay films are found to depths of 1.2 m in both the Deckard Flats deposits and in postglacial mass movement deposits probably less than a few thousand years old.

By Deckard Flats time, the icecap axis connecting Gallatin Range ice with that of the Washburn Range had lost its definition. The unrestricted movement of Deckard Flats ice eastward from the Gallatin Range and the southwestward flow at the northwest end of the Washburn Range indicate that the icecap between the Gallatin and Washburn Ranges had greatly diminished or entirely disappeared.

RECESSION FOLLOWING THE DECKARD FLATS READJUSTMENT

Glaciers from the Gallatin Range.—After Deckard Flats time, ice east of the Gallatin Range receded rapidly into individual valleys. Recessional deposits and erosional features marking this recession are scant and consist mainly of kame gravels at the margins of the main topographic lows and some melt-water channels and eskers in the middle of Gardners Hole (frontispiece).

Glaciers from the Beartooth uplift.—The best recessional features developed after the Deckard Flats readjustment are between Blacktail Deer Plateau and Junction Butte. The most prominent features are deep ice-marginal melt-water channels, such as the one which contains Phantom Lake. The old Mammoth-Tower road traverses an exceptionally well developed ice-marginal channel known as "the cut" (fig. 40). Although the channels are prominent, glacial moraines associated with them are minor, if present at all. The positions of kame terraces and melt-water channels indicate simultaneous channel cutting across bedrock salients and kame gravel deposition in reentrants. The channels were probably first eroded into bedrock during pre-Pinedale glaciations and flushed out and deepened during the Pinedale recession.

The southward flow of Deckard Flats ice into the Tower Creek amphitheater and Antelope Creek area seems to have reversed to northerly when the moraines 150–305 m below the Deckard Flats limit were deposited. Although abundant erratics of Precambrian rocks from the Beartooth uplift indicate a northern source, some of the lateral morainal ridges slope to the north rather than south. A possible explanation for this is a glacial surge to the south, followed by a lowering of ice levels in the Yellowstone valley and northward flow out of a basin here called the Tower Creek "amphitheater." The main reason for suggesting a glacial surge is that it provides a mechanism that could both push ice into the Tower Creek amphitheater and, before this ice melted, draw down the level of ice in the Yellowstone valley enough so that the

FIGURE 40.—Oblique view south along a melt-water channel known as "the cut." The channel transects ice-streaming, molding and scouring features (left to right lineation of topography) formed during full-glacial time. The channel is as much as 60 m deep and was occupied by the ice-marginal Yellowstone River as the glacier receded from the Deckard Flats position. Extensive deglaciation had occurred by the time ice-marginal melt waters flowed through "the cut", for the ice surface was more than 600 m higher when it striated Prospect Peak. (See also fig. 26.)

"stranded" ice could flow back out of the Tower Creek amphitheater. A glacial advance from the north followed by one from the southwest might also account for these features, but is considered less likely for it involves a separate advance for which no end moraines have been observed.

When the Beartooth ice margin had receded to the Tower Falls area, the Grand Canyon of the Yellowstone was ice free but was blocked at its lower end by the southwest margin of Beartooth ice (fig. 41). This dam formed a lake approximately 180 m deep at an altitude of 2,070 m (6,800 ft) called "Retreat Lake" by Howard (1937, p. 131–137). A stable outlet for the lake was provided by a melt-water channel around an unnamed knob and through a channel now occupied by Lost Lake. This channel is mostly just below 2,070 m and can be traced nearly continuously for 10 km. Retreat lake nearly filled with silt before the Beartooth ice receded. When the ice receded enough so that the Lost Lake channel was abandoned, lower ice marginal channels such as the one now occupied by the road between Tower Junction and Tower Falls (altitude of about 1,950 m or 6,400 ft) were occupied. As progressively lower outlets were used and "The Narrows" canyon was cut, the lake sediments that had nearly filled retreat lake were incised by the Yellowstone River, leaving several gravel-mantled terraces cut into these sediments.

AGE OF THE DECKARD FLATS READJUSTMENT

In terms of the standard Rocky Mountain glacial sequence (Richmond, 1965a), the Deckard Flats readjustment is probably of late middle Pinedale age. The Deckard Flats is older than carbonaceous deposits from the base of the Swan Lake depression dated at 13,530±130 years old (WIS-432). However, this age may be as much as 1,000 years too old due to incorporation of old carbon from nearby limestones. Two other carbon dates indicate that the Yellowstone plateau was mostly deglaciated by 13–14,000 years ago; ages of 14,360±400 years from the bottom of a bog (W-2780, Waddington and Wright, 1974) and 13,140±700 years from lake sediments (W-2037; Richmond and Pierce, 1972) were determined from samples collected 140 and 110 m respectively above Yellowstone Lake. Although no maximum dates have been obtained for the Deckard Flats, it is probably no older than the time of major deglaciation, dated between 15,000 and 11,000 years for mountain glaciations. Thus, the age of the Deckard Flats end moraines is estimated to be between 15,000 and 13,000 years old.

No differences in the degree of soil development were noted between the Pinedale terminal moraines and the deposits of Deckard Flats age. Thus the soils indicate no major relative-age difference.

On the other hand, weathering rinds on basalt cobbles from the B-soil horizon in Deckard Flats deposits are only one-fourth as thick as those from Pinedale terminal moraines in a similar weathering environment. Deckard Flats deposits on Swan Lake Flat have rinds that average 0.10±0.07 mm in thickness; Pinedale moraines near West Yellowstone have rinds that average 0.40±0.22 mm in thickness (Colman, 1977, p. 195). A conservative evaluation of this rind thickness data suggests that the Deckard Flats is less than 15,000 years old, or less than half the age of the Pinedale end moraines, which are dated by obsidian hydration as about 30,000 years old (Pierce and others, 1976).

Obsidian occurs in recessional Deckard Flats deposits along Obsidian Creek (Mammoth quadrangle). The thickness of hydration rinds indicates that they were glacially abraded 10,000 to 15,000 years ago (Pierce and others, 1976, fig. 4).

LATE PINEDALE ADVANCE OR STANDSTILL

End moraines located farthest upvalley, excluding Neoglacial cirque moraines, are classified as of late Pinedale age in Yellowstone Park. Soil profiles and other weathering characteristics of these deposits

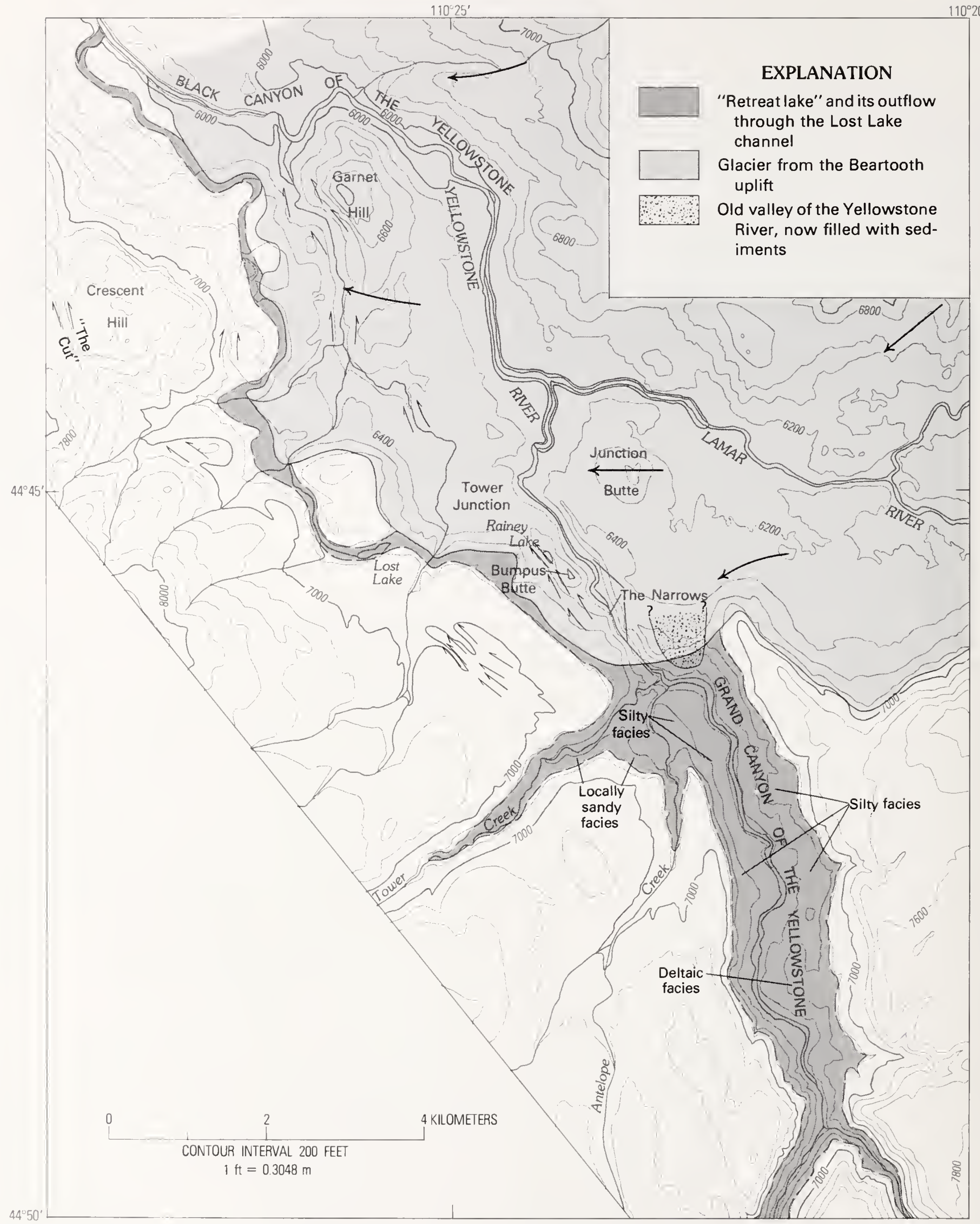


FIGURE 41.—"Retreat lake" in the lower Grand Canyon of the Yellowstone River dammed by ice from the Beartooth uplift. Retreat lake filled with about 180 m of sediments, which are mostly silty but coarsen where the Yellowstone River and Tower Creek entered the lake. The rock threshold at Lost Lake stabilized the level of the lake. Ice-marginal channels occupied both before and after Retreat lake drained through the Lost Lake channel are shown by half arrows. Larger arrows, flow direction of ice. Base from U.S. Geological Survey 1:62,500 Tower Junction, 1959.

are similar to the full-glacial and recessional Pinedale deposits (Richmond, 1968). Moraines of the late Pinedale are poorly expressed or absent downvalley from most poorly developed or southerly facing cirques.

GALLATIN RANGE GLACIERS

Three distinct, horseshoe-shaped end moraines about 10 m high occur on valley floors 1.5–10 km from the headwalls of well-developed cirques in the Gallatin Range (pl. 3; fig. 46; Pierce, 1973a). They represent an advance of local mountain-valley glaciers. An outwash terrace extends downstream from the outermost of these moraines in the valleys of the Gallatin River and Indian Creek, and probably also in the valleys of Panther and Fawn Creeks.

ICECAP ON THE BEARTOOTH UPLIFT

No clear-cut late Pinedale advance was recognized for Beartooth ice. No end moraines with well-defined outer limits occur between the cirques and Junction Butte, but, as noted by Alden (1928, p. 62), well-developed morainal topography is present in the Junction Butte area just east of Tower Junction (fig. 42; pl. 3). Strong morainal topography with abundant depressions and numerous large boulders of Precambrian rocks is evident from Junction Butte for about 6 km up the Lamar valley. The lack of a clearly defined outer limit for this morainal topography is inferred to signify that here the late Pinedale ice only paused in its overall recession.

RECESSION FROM THE LATE PINEDALE POSITION

GALLATIN RANGE GLACIERS

Several moraines were deposited during pauses in the recession of glaciers from their late Pinedale stands. Along the Gallatin River, three sets of low moraines are rather evenly spaced between the end moraines of late Pinedale age and the valley head. Along the east side of the range, several recessional moraines were deposited upstream from the late Pinedale terminus in flat-bottomed valleys (Pierce, 1973a).

GLACIERS FROM THE BEARTOOTH UPLIFT

During recession of the late Pinedale Beartooth icecap from its Junction Butte position, trunk glaciers in the valleys of Slough Creek and Soda Butte Creek continued to flow while the glaciers from cirques in tributary valleys stagnated and disappeared. As described in the following sections, these changes provide evidence of phase relations similar to those already described between the northern Yellowstone outlet glacier and the local tributary glaciers.

Slough Creek glacier.—Large erratics of Precambrian rocks, along with lateral moraines whose crests slope up the Lamar valley, indicate that a glacier flowing down the valley of Slough Creek advanced up the valley of the Lamar River, probably after the late Pinedale maximum. These features are best displayed in the Jasper Creek area just upstream from Lamar Canyon. An abundance of large subangular boulders of Precambrian rocks in moraines above Lamar Canyon was observed by John Good (written commun., 1972), who noted the similarity in general appearance between these moraines and those near Junction Butte. I concur with his conclusion that the source of the erratics is the lower reaches of Slough Creek only 3 km distant, rather than the Beartooth Mountains east of Cooke City 41 km away (pl. 3). The fact that erratics become much less common up the Lamar and Soda Butte valleys, toward any Beartooth source, supports this conclusion.

The Lamar valley is flat bottomed and about 1.5 km wide for 10 km upstream from Lamar Canyon; almost no bedrock is exposed in this area. In full-glacial times, ice scoured a rock basin in the readily erodible Eocene volcanics upvalley from the resistant Precambrian rocks in Lamar Canyon. This rock basin is now filled with an unknown thickness of glacial sediments.

Lake silts and sands on the downstream side of Lamar Canyon (Howard, 1937, p. 145) indicate that as the Slough Creek glacier receded, the ice dam it had formed in the Lamar River valley receded to below Lamar Canyon. These lake sediments are poorly exposed, for they are mantled by a veneer of gravel (Pierce, 1974a). The sudden release of temporary lakes formed behind this dam are thought to be the cause of catastropic floods along the Yellowstone-Lamar drainage. where Slough Creek enters the Lamar valley, a nested set of kame terraces (fig. 43) indicates progressively lower ice levels as the Slough Creek glacier receded.

Soda Butte glacier.—The glacier in Soda Butte Creek probably joined the Slough Creek glacier in late Pinedale time and then separated from it sometime during late Pinedale recession. No end moraines were noted between Cooke City and the moraines of late Pinedale age near Junction Butte. The main Soda Butte glacier was augmented by glaciers from local, well-developed cirques. The glacier from the icecap on the Beartooth uplift continued to flow down the valley of Soda Butte Creek longer than the tributary glaciers from either Pebble or Amphitheater Creeks. Kame deposits more than 30 m thick occur in the lower parts of these two valleys (Pierce, 1974b).

The distribution of erratics also shows that ice from the Beartooth icecap outlasted glaciers from local cirques in the Cooke City area. Erratics of Precambrian

FIGURE 42.—Pond and large erratics in moraines of late Pinedale age near Junction Butte. View is to the southeast; glacier source mostly out of sight in the left distance. Photograph by W. R. Keefer.

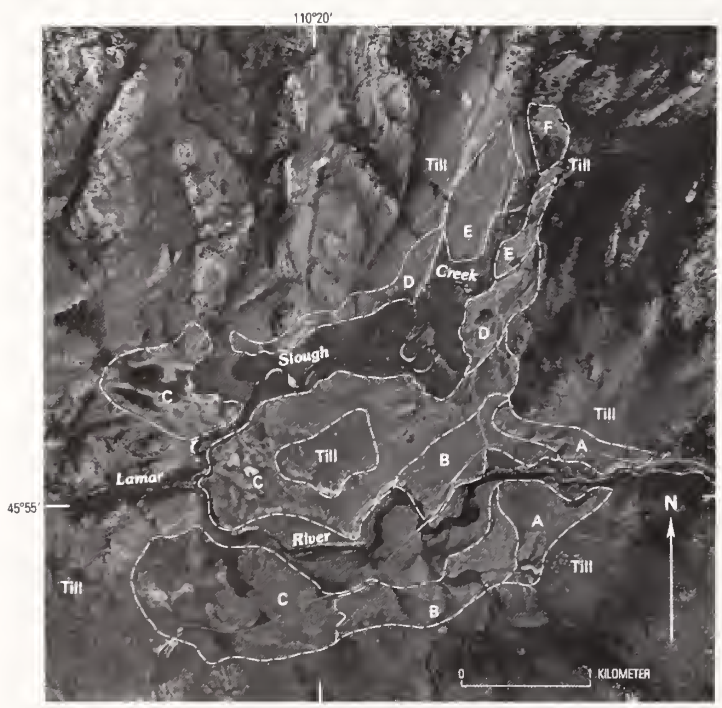


FIGURE 43.—Sequence of six nested kame deposits from oldest (A) to youngest (F) at mouth of Slough Creek. These deposits consist of ice-contact stratified drift of gravel and sand. Note numerous kettles and kettle-lakes. The glacier flowing southward along Slough Creek blocked the Lamar River during deposition of units A, B, and C; lake sediments veneered by gravel show that ice-dammed lakes were present. Photograph by U.S. Geological Survey, 1954.

crystalline rock carried by the Soda Butte glacier occur as much as 1.5 km up Amphitheater Creek (pl. 3). Three kilometers west-northwest of the Northeast Entrance, erratics of Precambrian rocks were carried from Soda Butte Creek over the low divide into the Pebble Creek drainage (pl. 3). On the floors of several of the high valleys south of Cooke City, erratics of Precambrian rocks occur in areas once occupied by late Pinedale glaciers from the local cirques. No Precambrian bedrock occurs in the local cirques, however, so the only likely source for them is the Beartooth uplift northeast of Cooke City. For example, just outside the local cirque moraines between Abiathar and Amphitheater Peaks, is found a Precambrian granitic erratic 1 m in diameter (Pierce, 1974b). This suggests that the Soda Butte glacier was 400 m thick when the late Pinedale glacier from this large, high local cirque was less than 0.7 km long. Similar relations exist in high valleys to the east. For example, in the valley of Republic Creek, W. G. Pierce (written commun., 1972) reported erratics of Precambrian rocks to altitudes of 2,500 m (8,200 ft), 220 m above Soda Butte Creek; and in the valley of Hayden Creek, he reported similar erratics to altitudes of 2,600 m (8,600 ft), 250 m above Soda Butte Creek.

AGE LIMITS OF THE LATE PINEDALE ADVANCE OR STANDSTILL

The radiometric age of glacial deposits of late Pinedale age in northern Yellowstone Park is known only within broad limits. On the east side of the Gallatin Range, several miles outside the late Pinedale limit, organic material at the base of the broad Swan Lake depression has a ^{14}C age of 13,530±130 years

(WIS-432). It should be noted that this age may be as much as 1,000 years too old due to "dead" carbon from nearby limestones being taken up by plants growing underwater. This sample is several miles outside end moraines of late Pinedale age and seems to closely postdate deglaciation. Thus, the late Pinedale is probably younger than 13,500 years old and perhaps younger than 12,500 years old.

In the Eagle Peak quadrangle (Richmond and Pierce, 1972, section 14), gravel of inferred late Pinedale age overlies peaty material collected by the author and dated as 13,140±700 years B.P. (sample W-2037; Meyer Rubin, written commun., 1969).

An age for the late Pinedale no younger than about 11,000 years is indicated by the climate inferred from the pollen record, which suggests that from about 11,000 to 5,000 years ago the climate was no cooler than that of the Neoglacial during the last 5,000 years (Baker, 1970, 1976; Bright, 1966; Waddington and Wright, 1974). Because late Pinedale glaciers were much larger than Neoglacial ones, the late Pinedale climate was probably at least as cool as the Neoglacial. Thus these bracketing ages suggest that the late Pinedale maximum occurred between 13,000 and 11,000 years ago.

Another aspect of the age of the late Pinedale is its relation to the major warming and deglaciation indicated in deep-sea cores (Broecker and others, 1960) that occurred about 11,000 years ago. It seems likely that it is older than the warming and deglaciation and is probably related to oscillations noted on various climatic curves in the 13,000–10,500-year interval.

LATE PINEDALE AND YOUNGER FLOODS

The occurrence of catastrophic floods during the Pleistocene (Bretz and others, 1956; Malde, 1968) has only recently been demonstrated to the general satisfaction of many geologists. Criteria developed to indicate the occurrence of these floods include giant flood bars, giant ripples, scablands, fluvial transport of huge boulders, and percussion-split and -spalled boulders. The hydraulic geometry along the flood path must be considered in order to understand the distribution of scour features, flood bars, and deposits of fine sediment. Scour occurs in valley constrictions, bouldery bars are deposited downstream from constrictions and from valley spurs, and fine sediment is deposited in areas of hydraulic damming upvalley from constrictions.

At least two large floods with waters 45–60 m deep rushed down the Lamar-Yellowstone drainage in late Pinedale time. The most impressive evidence of flooding can be seen 5 km northwest of Gardiner, along the south side of the Yellowstone River within 0.65 km of the park boundary. There between the road and the river is a midchannel bar 20 m high and 450 m across, whose surface is covered by giant current ripples (fig. 44). Arcuate ridges about 2 m high and 15 m apart trend across the bar. Ridgecrests are covered with boulders as much as 2 m in diameter (fig. 44). The ridges are slightly asymmetric and boulders are concentrated on the downstream sides of their crests (fig. 44). A massive longitudinal flood bar (figs. 44, 45) occurs to the southwest at the same altitude as the midchannel bar. Other elongate bouldery ridges parallel to the valley axis can be traced for as much as a

FIGURE 44.—Giant current ripples on midchannel flood bar north of Gardiner, Mont. *A*, Oblique aerial photograph showing form of bar and pattern of giant current ripples. Flood bar is about 500 m across and 20 m high. Crests of ripples (fig. 45) are about 2 m high and about 15 m apart. Flow was towards the camera. A longitudinal flood bar is shown in the upper right of the photograph. Photograph by J. S. Shelton, 1970. *B*, Ridges of boulders forming giant current ripples on the midchannel flood bar. The boulders are concentrated on the downstream (right) side of the crest.

kilometer; the midchannel flood bar is about 1.5 km long. The material in this flood bar was probably largely reworked from Deckard Flats end moraines.

The best evidence for both the height of flooding and the occurrence of multiple floods is at the truncated fronts of alluvial fans at the mouth of Reese Creek, along the park boundary (fig. 45). The fans are cut by abrupt east-facing escarpments 3–15 m high that parallel the Yellowstone River. On and below the escarpments, evidence of flooding exists in the form of longitudinal bars, ranging in length from 3 to 300 m. As shown in figure 45, at least two floods, the first reaching 60 m and the second reaching 45 m above the present Yellowstone River, are indicated by the configurations of flood-ripped fan fronts.

Upstream from the Reese Creek fans, flood deposits are well displayed on the south side of the Yellowstone River near Gardiner. Three kilometers northwest of Gardiner, the downstream end of a long, cigar-shaped flood bar emerges from beneath a Holocene earthflow. Gardiner High School is built partly on a midchannel bar, whose bouldery crest resembles a Pinedale moraine. However, the concave downvalley trend of some of the crest lines on the bar suggest that it is a flood feature. The interior of this flood bar and another bar about half a kilometer downvalley are well exposed in railroad cuts. Although boulders are concentrated on the bar surfaces, their interiors locally are mostly sand. The flood deposits are moderately well sorted. Boulders of fine-grained rocks such as basalt often show percussion spalls, especially on angular corners, due to collisions of boulders during flood transport.

Along the Yellowstone River near the mouth of Bear Creek, a jet of floodwater from the Black Canyon of the Yellowstone River eroded bedrock into scablands and left bouldery flood deposits to heights of about 60 m above the present river.

Just west of the north entrance station to the park, floodwaters appear to have ripped the upper surface of a gravel terrace and eroded a scarp into it in a manner similar to that of the fans at the mouth of Reese Creek. The terrace is underlain by glacial gravels derived mostly from the Mammoth Hot Springs–Gallatin Range area. The northern end of this bench is a flood-eroded escarpment mantled by small cigar-shaped bars that are littered with abundant boulders of Precambrian rocks from the Black Canyon of the Yellowstone.

In the Black Canyon of the Yellowstone, flood features are not as obvious as in the Gardiner area. On the south side of the river, just upstream from Crevice Lake, flood deposits surround a low knob of bedrock with a scabland appearance. The origin of the depression now occupied by Crevice Lake (pl. 3; Pierce, 1973b) is an enigma. Since the deposits enclosing the lake are streamlined and appear to be flood gravels, the lake may be either an ice-block depression in flood debris, or an erosional feature formed by a flood vortex caused by the configuration of the canyon.

Farther upstream, near the confluence of the Lamar and Yellowstone Rivers, flood deposits accumulated in the wider parts of the valley, both in deceleration areas below constrictions and in hydraulically dammed areas above the constrictions (fig. 46).

Erosional and depositional features caused by floods occur along the Lamar valley at least as far upstream as Slough Creek (pl. 3). Deposits apparently formed by floods continue upstream as far as Lamar Canyon, but they are not as well expressed nor as high above present drainage as those downstream from Slough Creek. The paucity of flood features in this area may result from the presence of the snout of the Slough Creek glacier, over which the flood waters may have debouched. As mentioned previously, the presence of glacially dammed lakes is indicated by lake sediments along Lamar valley above Slough Creek. Catastrophic release of these ice-dammed lakes appears to have produced the largest flood flows down the Lamar-Yellowstone drainage. Howard (1937, p. 144–145) anticipated the interpretation presented here by concluding that "Lamar lake" was dammed by the Slough Creek glacier and "may have drained rapidly."

Flood deposits along the Yellowstone River above its confluence with the Lamar indicate that waters about 25 m deep coursed through The Narrows some time after the deepest floods went down the Lamar-Yellowstone drainage. Flood deposits about 15 m above the river on the point at the Yellowstone-Lamar confluence contain abundant percussion-spalled boulders of basalt. The topograpic form of the deposits indicates flow down the Yellowstone. Also, the only reasonable source for the basalt is along the Yellowstone in The Narrows area; no concentrations of basalt are present for many kilometers up the Lamar. The location of this deposit in the path of the larger floods down the Lamar but well below their upper limit indicates that this flood was younger and smaller than those that came down the Lamar.

Floodwaters debouching from The Narrows of the Yellowstone formed a flood bar just upstream from the highway bridge across the Yellowstone River. This bar extends from a spur and forms a ridge separated from the valley wall by a dry trough about 7 m deep. Although the surface of the bar is rather smooth and not remarkably bouldery, exposures in the bank of the Yellowstone River at the upstream end of the bar reveal abundant boulders more than 1.6 m across, and

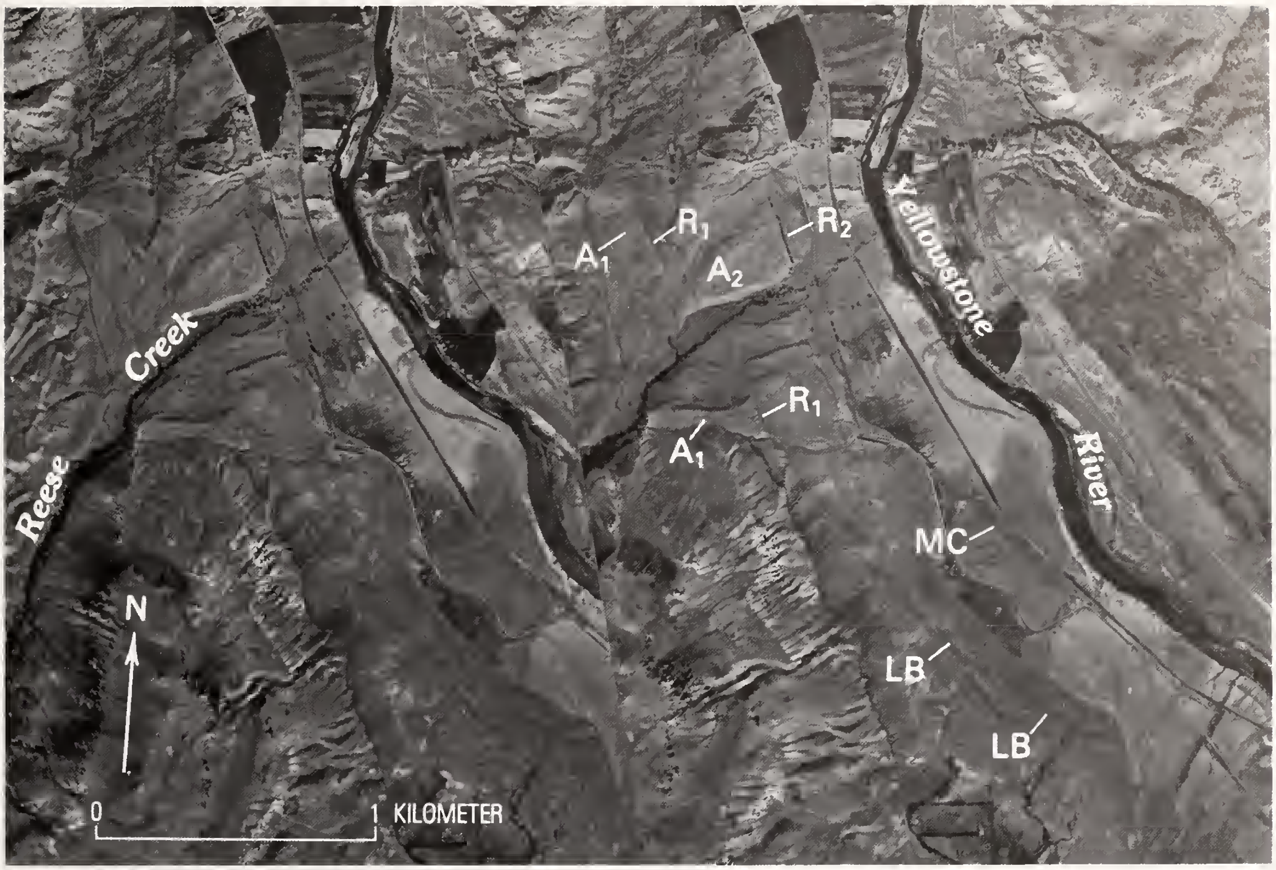


FIGURE 45.—Stereopair of aerial photographs showing evidence of two flood episodes along the Yellowstone River near Reese Creek. The inferred sequence of events is: (1) alluvial fan (A_1) was built by Reese Creek; (2) flood of water about 60 m deep eroded much of that fan and left a flood-ripped fan front (R_1) close to the valley wall; (3) younger alluvial fan (A_2) was built by Reese Creek; and (4) younger flood of water about 50 m deep eroded that fan and left a flood-ripped fan front (R_2) with longitudinal flow features along and below the fan front. The height of the second flood is indicated by modification of the fan surface up to about 10 m above the top of the fan front. The midchannel flood bar of figure 44 (MC) can be seen 1.3 km southeast of Reese Creek, flanked by a longitudinal flood bar (LB) to the south and west. Aerial photographs by U.S. Geological Survey, 1954.

one boulder of locally derived basalt 10 m across.

The upstream source of this flood is unknown. No definite flood deposits were recognized above The Narrows, but hydraulic damming by The Narrows would reduce upstream flood velocities and make it difficult to separate flood deposits from the deposits of "Retreat lake," described earlier. Inasmuch as this flood occurred after the late Pinedale floods down the Lamar-Yellowstone drainage, it probably is not related to ice from the Beartooth uplift. It seems likely that the flood resulted from failure of a landslide dam either in The Narrows area or in the Grand Canyon of the Yellowstone River upstream from the main area of "Retreat lake" sediments.

Evidence of floodwaters from the park boundary down the Yellowstone to Yankee Jim Canyon consists mainly of longitudinal bars, such as the one at the base of Cinnabar Mountain directly across from Corwin Springs. This bar, about 1.5 km long, resembles a terrace, but its streamlined form and the absence of an undercut scarp at its outer edge show that it is not a normal stream terrace.

The history of late Pinedale flooding from Yankee Jim Canyon downvalley is complicated by the younger Yankee Jim flood (pl. 3). Good (1965) recognized that a large landslide on the west side of Dome Mountain had dammed the Yellowstone River to a depth of about 55 m. Presumably, when the landslide dam was overtopped, rapid erosion of the spillway released the Yankee Jim flood. Immediately below the landslide dam and 15 m above the river, a large flood deposit contains huge blocks of locally derived Precambrian bedrock. Channel morphology shown on aerial photographs suggests that this deposit is a fan-like accumulation derived from the landslide.

Downvalley from Yankee Jim Canyon the Yankee

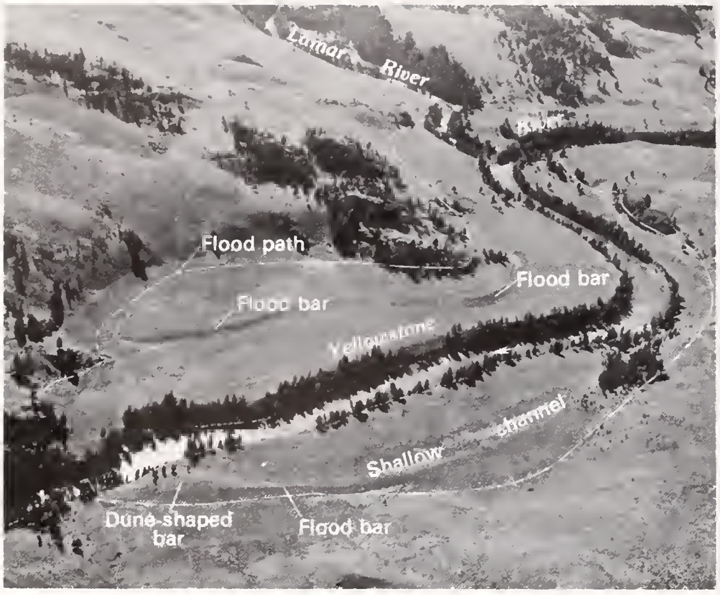


FIGURE 46.—Oblique aerial view of flood deposits below the junction of Lamar River with the Yellowstone. Floodwaters more than 50 m deep from the Lamar River left deposits in the wider parts of the Yellowstone valley. A bedrock constriction at the downstream edge of the photograph caused hydraulic damming of the floodwater, resulting in lower velocities and deposition of floodgravels, especially against the valley walls and downstream from spurs. A shallow channel on the flood deposits parallels the near side of the Yellowstone River. Photograph by W. B. Hall, 1970.

Jim flood seems to have reached about half as high as the highest late Pinedale flood. The fan-like flood deposit immediately below the Dome Mountain landslide would have been obliterated or modified by the higher floodwaters of the Lamar floods, so the Yankee Jim flood is dated as younger than the main floods down the Lamar-Yellowstone drainage. At the downstream end of Yankee Jim Canyon, flood deposits, probably formed by the Yankee Jim flood, underlie the boulder-strewn flats 15 m above the Yellowstone.

Longitudinal flood bars 60 m above the Yellowstone River occur downvalley from the spur at the northeastern end of the canyon. They are probably related to the largest floods that came down the Lamar-Yellowstone drainage.

In the Chico end moraine area 4 km downstream from Emigrant (pl. 2), stony deposits in the "inner valley" of the Yellowstone were once thought to represent terminal moraines of the late Wisconsin (Pinedale) (Alden, 1932, p. 123; Horberg, 1940, p. 297). These deposits are within the area covered by floods and appear related primarily to them. The deposits are hummocky on a small scale but have a streamlined form parallel to the valley axis and locally slope upvalley. On the east side of the valley, they grade into a surface that lies about 10 m above the river and resembles a giant flood bar. This alluvial surface terminates downstream in a delta-like front perpendicular to the direction of flow, indicating that it is not a normal stream terrace.

GLACIAL DYNAMICS INFERRED FROM GEOLOGIC STUDIES

This final section concerns subjects derivative from the field reconstruction of the late Pleistocene glaciations. It deals with three main topics: (1) the differences in the relative extent of the Bull Lake and Pinedale icemasses north and west of the park and the reasons for these differences, (2) quantitative evaluations of the Pinedale reconstruction (pl. 1*A*) by calculation of basal shear stresses and mass balance, and (3) the phase relations between the different kinds of glacial source areas throughout a glacial cycle.

COMPARISON OF BULL LAKE AND PINEDALE GLACIATIONS

In the West Yellowstone area, Bull Lake glaciers from Yellowstone Park extended about 20 km west beyond the Pinedale ones (fig. 9). Southwest of the park, G. M. Richmond (1976, fig. 7) mapped the outer boundary of moraines of possible Bull Lake age about 20 km beyond those of Pinedale age. However, the outlet glacier flowing north from Yellowstone was larger in Pinedale time than during the Bull Lake Glaciation. A Pinedale age is indicated for both the Eightmile and Chico end moraines by their morphology, soils, boulder weathering, morainal backfills, ice-dammed lake sediments, and relations to outwash deposits. Nearby moraines of the western Mount Cowen glaciers (fig. 12) provide examples of the contrast between Bull Lake and Pinedale moraines and the nearly equivalent size of local Bull Lake and Pinedale glaciers. This similarity suggests that a minor increase in the size or change in the flow direction of Pinedale glaciers could easily result in obliteration of Bull Lake moraines.

Thus, although the Bull Lake Glaciation was more extensive than the Pinedale on the west side of the park, it was less extensive on the north (fig. 47). The most likely reason for this difference is the eruption of large rhyolite flows after the Bull Lake and before the Pinedale glacial maxima (fig. 47).

Southward from Madison Canyon, a continuous string of rhyolite flows all have K-Ar ages indicating their emplacement between the Pinedale and Bull Lake glacial maxima. According to combined obsidian-hydration and K-Ar dating, the Bull Lake glacial max-

imum near West Yellowstone occurred about 150,000 years ago and the Pinedale about 30,000 years ago (Pierce and others, 1976). For rhyolite flows from Madison Canyon southward, John Obradovich (written commun., 1979) reported the following K-Ar ages (in years): West Yellowstone flow, 117,000±8,000; Summit Lake flow, 110,000±25,000; and Bechler River flow, 117,000±7,000 (all weighted means of multiple runs, and weighted standard deviation, $2\bar{\sigma}_x$); and Pitchstone Plateau flow, 70,000±2,200 (1σ, instrumental error only). Field relations independently show that the West Yellowstone flow was emplaced between the Pinedale and Bull Lake glacial maxima.

Emplacement of these flows resulted in a major topographic change, adding a constructional volcanic landform that averages about 150 m in thickness and extended from Madison Canyon 55 km southward to the Pitchstone Plateau. This volcanic landform acted as a diversion dam to flow of Pinedale ice, restricting westward flow relative to flow during the Bull Lake glaciation, and increasing flow to the north.

Between Pinedale and Bull Lake time, the eruption of several lava flows increased the altitude of the plateau surface just southwest of the Washburn Range (fig. 47). This increase in altitude favored development of a larger Pinedale icecap in this area, adding to the size of the northern Yellowstone glacier as well as the glaciers near the southern end of the Gallatin Range.

QUANTITATIVE EVALUATION OF THE RECONSTRUCTED ICEMASS

The reconstruction of the full-glacial icemass of Pinedale age (pl. 1*A*) is based on field observations. Glaciology offers methods for independent evaluation of this reconstruction; the only assumption necessary is that former glaciers obeyed the same physical laws and were affected by the same physical processes as modern glaciers.

BASAL SHEAR STRESS

Calculation of basal shear stress is a powerful tool in evaluating the plausibility of reconstructions of former icemasses. The two parameters that together specify the shape of a reconstructed glacier—ice thickness and surface slope—are also the primary variables that determine basal shear stress.

The formula for determining basal shear stress of an icecap (Nye, 1952) is $\tau_b=\varrho gh \sin\alpha$. Where τ_b=basal shear stress in bars, ϱ=density, G=acceleration of gravity, h=the ice thickness, and α=slope of the ice surface. For the vast majority of modern glaciers, empirical observations indicate that the basal shear stress is between 0.5 and 1.5 bars and tends to be constant for any relatively homogeneous segment of an icecap or valley glacier (Nye, 1952; Paterson, 1969, p. 91). In addition, calculated basal shear stress for well-studied Pleistocene glaciers also generally[4] falls within these limits (Mathews, 1967; Clark, 1967).

Some reasons for the narrow range of values for basal shear stress are well stated by Mathews (1967, p. 151):

> From laboratory tests it has been inferred that if the thickness or surface slope of an ice sheet somehow became so great that basal shear stress greatly exceeded this fixed value, the basal ice could be deformed so easily that the sheet would very rapidly spread out until its thickness or slope was significantly reduced. Conversely, if because of combination of thinness and flatness in an ice sheet the basal shear stress was very low, movement would virtually cease until accumulation of more ice increased thickness or slopes or both, and these in turn increased the basal shear stress.

The comparison of glacial flow rates to the variables that determine basal shear stress explains in part why the values of basal shear stress tend to vary within narrow limits. Rates of flow and glacial discharge are functions to the cube or higher power of the same variables that determine basal shear stress. In laboratory experiments, Glen (1955) determined that the effective shear strain rate ($\dot{\nu}$) of polycrystalline ice is proportional to the power n of the applied stress τ, or $\dot{\nu}\propto\tau^n$. For stresses between 0.8 and 5.5 bars, the exponent n is between 2 and 4.5; some observations on modern glaciers suggest a value of 3 for the exponent n is appropriate (Nye, 1965, p. 662). The volume of ice passing through a given cross section in unit time is a function of the slope α to the power n and of the thickness h to the power $n+2$ (Nye, 1952, p. 84). Assuming a value of 3 for n, a doubling of either the slope or thickness would only double basal shear stress, but would increase discharge by a factor of 8 or 32, respectively.

A shape factor F modifies the equation for basal shear stress when glacial flow is affected by drag against valley walls (Nye, 1965; table 3, this report). As shown by Nye (1965), this shape-factor depends primarily on W, the ratio of the half-width of the glacier to its maximum thickness, and to a lesser extent on the shape of the valley cross section. The

[4]Mathews (1974) determined values of basal shear stress below 0.2 bars for some lobes of the Laurentide Ice Sheet. He noted (p. 41) that the anomalously low values of basal shear stress were associated with the ice lobes on terrain with low relief, and that the low basal shear stress may be related to water confined under high hydrostatic pressure at the base of the glacier. All the areas for which he calculated anomalously low basal shear stresses are underlain by impermeable shale. The shale would have acted as a seal that prevented escape of water from beneath the glacier, thereby permitting fluid pressures approaching that of the "icesostatic" pressure of the overriding ice. Compared to the sites studied by Mathews, the northern Yellowstone area differs because it has much more rugged topography, much narrower glaciers, and no extensive terrane underlain by impermeable shale.

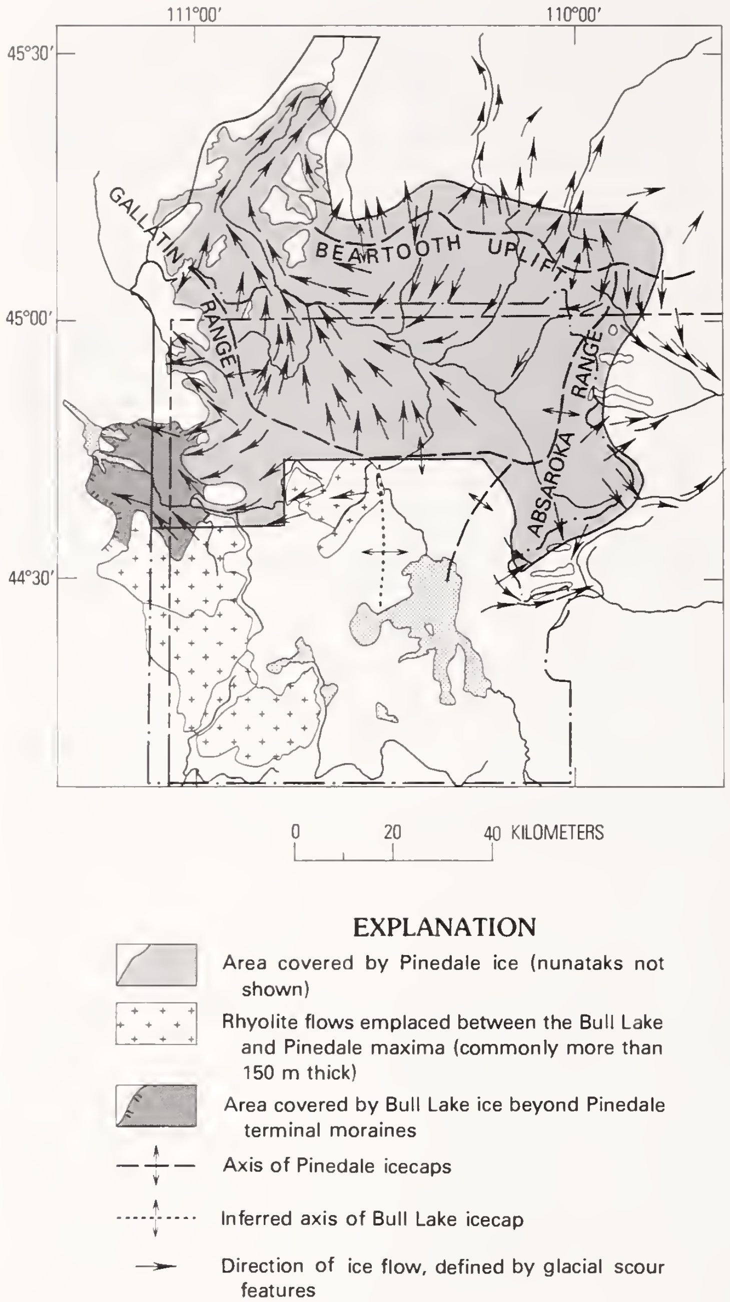


FIGURE 47.—Comparison of Bull Lake and Pinedale icemasses in the northern Yellowstone area. West of the park, Bull Lake moraines extend as much as 20 km beyond those of Pinedale age. Between the Bull Lake and Pinedale maxima, large rhyolite flows were emplaced in the central and western part of the park. The rhyolite flows blocked westward flow of Pinedale ice and diverted it to the north. Pinedale ice overran Bull Lake terminal moraines in the valley of the Yellowstone River.

TABLE 3.—*Shape factor* (F) *for parabolic valley profiles*

[Shape factors for *W* from 1 to 4 are from Nye (1965, table 4); shape factors for *W* greater than 4 are extrapolated from Nye]

W	*F*	*W*	*F*
1	0.45	6	0.87
2	.65	8	.90
3	.75	16	.95
4	.81	∞	1.00

glaciated valleys in Yellowstone have parabolic cross sections and *W* was between 3 and 16, resulting in *F* values between 0.75 and 0.95 (table 3).

Thus, for glaciers flowing in valleys, the equation for basal shear stress is modified to:

$$\tau_b = F \cdot \varrho g h \sin\alpha$$

For the purposes of this paper, the result of the above formula is called primary basal shear stress. With uniform flow, primary basal shear stress and total basal shear stress are the same, but with strongly extending or compressing flow, longitudinal or transverse normal stress gradients can make significant contributions to the total shear stress. The calculation of these generally secondary factors is beyond the scope of this section, but discussions of the differential equations involved are given in Nye (1959, p. 505–506), Paterson (1969, p. 102–104), and Meier (1968).

Isostatic deformation due to glacial loading would have depressed the central glaciated area of northern Yellowstone relative to the glacial margins. If we use the isostatic deformation of Lake Bonneville as a model (Crittenden, 1963), about 100 m of central depression is possible for ice averaging 600 m in thickness. This is about 5 percent of the altitude difference from the terminus to the ice divides of the northern Yellowstone glacier. This isostatic effect, not accounted for in the glacial reconstruction, would reduce reconstructed values for glacial slopes and calculated basal shear stress by only about 5 percent.

Figure 48 shows a simple way of both calculating and graphically displaying the relations between ice thickness, surface slope, and primary basal shear stress. The design of this plot is modified from that made for rock glaciers by Wahrhaftig and Cox (1959, fig. 10). In figure 48, calculated basal shear stresses are shown as straight lines. The equation for this relation between effective thickness and the cosecant α is derived by deleting the values ϱ and g which are constant, thereby giving:

$$\tau_b \propto \sin\alpha \cdot F \cdot h$$

and then converting sin α to 1/cosecant α giving:

$$\tau_b \propto \frac{F \cdot h}{\text{cosecant } \alpha}$$

The value of the numerator of the last equation ($F \cdot h$) is here called the "effective thickness." The last equation illustrates that a constant value of primary basal shear stress will plot as a straight line when the effective thickness is plotted against cosecant α.

Figure 48 graphically shows the primary basal shear stress of 50 reaches covering most of the northern Yellowstone area. Each reach is between 5 and 15 km long and was selected to have a consistent flow pattern throughout its length. (Location and length of reaches

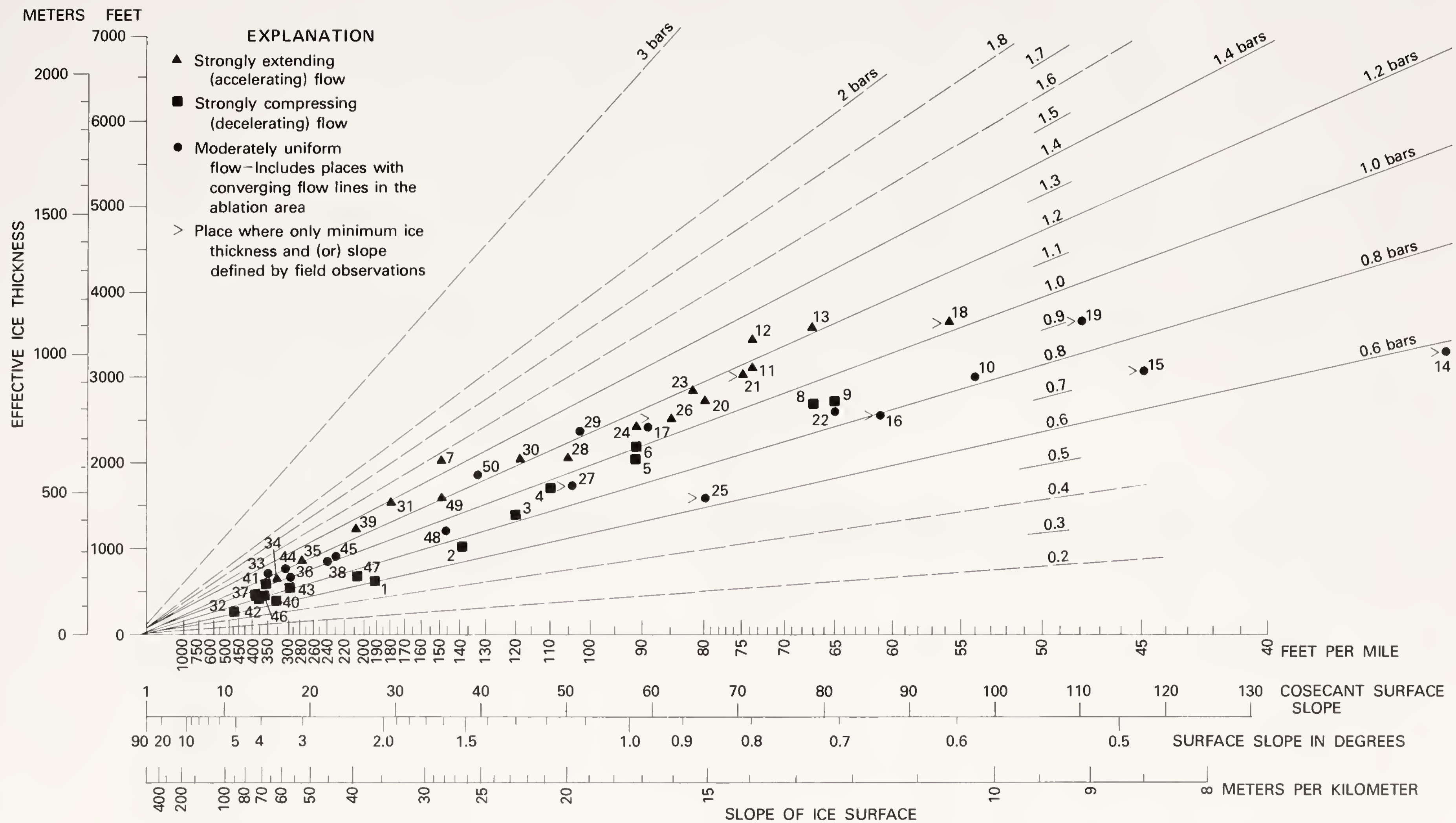


FIGURE 48.—Values of effective ice thickness, slope, and primary basal shear stress for the Pinedale full-glacial icemass (pl. 1*A*). All values of primary basal shear stress plot between the rather narrow limits of 0.6 and 1.5 bars. The 50 sites represent reaches 5–15 km long, which end-to-end cover the main flow paths of the glacial system in northern Yellowstone. The numerical values for each reach are given in table 4.

are given in table 4.) This use of relatively long reaches apparently eliminated most of the local variations in primary basal shear stress calculated for shorter reaches (Clark, 1967; Meier, 1960, p. 49; Meier and others, 1974, p. 210). Because the reaches are mostly end to end, the calculations simulate an integral calculation for the icemass. The slopes or thicknesses of one reach cannot be changed without affecting the basal shear stress of contiguous reaches.

The values of primary basal shear stress fall between 0.6 and 1.5 bars (fig. 48; table 4), consistent with modern glaciers where "the great majority of values are between 0.5 and 1.5 bars" (Patterson, 1969, p. 91).

Reaches of a glacier are classified on the basis of their flow pattern. If the velocity increases downstream, the flow is extending (or accelerating with distance downstream). The following conditions promote extending flow: (1) downflow convergence of glacier margins in plan view, (2) downflow thinning over bedrock prominences (convex longitudinal valley profile), and (3) addition of ice in the accumulation area.

If the velocity decreases downstream, the flow is compressing (or decelerating with distance downstream). The following conditions promote compressing flow: (1) downflow divergence of glacier margins in plan view, (2) downflow increase in ice thickness (concave longitudinal valley profile), and (3) removal of ice in the ablation area.

As shown by the different symbols in figure 48, higher values of primary basal shear stress occur in areas of extending (accelerating) flow, such as the northward convergence to form the northern Yellowstone outlet glacier (fig. 21). Lower values of primary basal shear stress occur in areas of compressing (decelerating) flow, which are located mostly below the 2,590 m (8,500 ft) contour in the ablation area. The 50 reaches that together cover all the main flow paths of the total icemass in the northern Yellowstone area (pl. 1*A*) are classified into reaches with strongly extending flow, reaches with nearly uniform flow, and reaches with strongly compressing flow (fig. 48; table 4). The primary basal shear stress for 15 strongly extending (accelerating) reaches averaged 1.21 ± 0.12 bars; the 11 nearly uniform reaches averaged 1.04 ± 0.16 bars; and the 16 strongly compressing (decelerating) averaged 0.84 ± 0.14 bars. (Eight reaches with only minimum depth values were not included in these calculations.) Hence, the variation in primary basal shear stress is strongly related to flow pattern, especially to convergence in plan view, and to ablation.

The reconstructed glacial system shown on plate 1*A* has values of primary basal shear stress that are both internally consistent, and consistent with the values of primary basal shear stress of modern glaciers. This glaciological evaluation supports the validity of the reconstruction based on the glacial geologic studies.

BASAL SHEAR STRESS AND ALTERNATE GLACIAL RECONSTRUCTIONS

On the Surficial Geologic Map of Yellowstone National Park (U.S. Geol. Survey, 1972a), two alternatives are shown or implied for the size of the Pinedale icemass. Major differences in the height or extent of Pinedale ice are shown in the areas of the Washburn Range, Specimen Ridge, and south of the Gallatin Range. For reasons described here and earlier in this report, I conclude that Pinedale ice formed the fresh striations on the crests of the Washburn Range and Specimen Ridge when it flowed over these summits which reach altitudes of 2,925 to 3,122 m (9,600 to 10,243 ft). The Surficial Geologic Map of Yellowstone National Park shows a boundary near the base of the Washburn Range labeled "PL," for "maximum * * * extent of early Pinedale ice." This "Pinedale limit"is 450–700 m lower than the limit of Pinedale ice I have determined (pl. 1*A*). This lesser value for ice thickness in the source area on the Yellowstone plateau would reduce both the glacial thickness and surface slope to about half the values determined from the reconstruction shown on plate 1*A*. Calculations of values of primary basal shear stress for this "Pinedale limit" yield basal shear stress values between 0.05 and 0.15 bars. Calculations of values of primary basal shear stress based on the boundary marked "PL" on the Surficial Geologic Map of Yellowstone National Park in other areas in northern Yellowstone, such as the area west of the Washburn Range and the area south of the Gallatin Range, result in values between 0.0 and 0.1 bars. The lower option for the altitude of Pinedale ice (U.S. Geological Survey, 1972a) yields unreasonably low values of primary basal shear stress. Moreover, as argued earlier, the field relations also indicate that the lower option is not the Pinedale limit but represents phases in the recession of Pinedale ice.

TABLE 4.—*Values for basal shear calculations*

[Values given in English units because slope and thickness of ice (pl. 1) are from contours that are in English units]

No.	Midpoint of reach	Thickness h (ft)	Half width (ft)	W	f	Drop in feet / Distance in miles	Gradient (ft/mi)	Slope (α) in degrees	Effective thickness (f·h)	Basal shear stress	Flow type[3]
1	Glacier terminus-------	650	18,500	28	0.98	500/ 2.6	192	2.086	644	0.62	C
2	Emigrant---------------	1,100	14,500	13	.93	500/ 3.6	139	1.507	1,023	.72	C
3	Sixmile Creek----------	1,500	18,500	12	.93	500/ 4.15	120	1.307	1,395	.85	C
4	Big Creek--------------	1,900	18,500	10	.92	400/ 3.7	108	1.173	1,748	.95	C
5	Donahue Creek----------	2,300	7,000	8	.90	300/ 3.3	91	.986	2,070	.95	C
6	Tom Miner Creek--------	2,550	14,000	5.7	.86	300/ 3.3	91	.986	2,193	1.00	C
7	Yankee Jim Canyon------	2,750	8,000	2.9	.74	300/ 2.0	150	1.628	2,035	1.54	E
8	Slip and Slide Creek---	3,200	17,000	5.3	.85	300/ 4.5	67	.723	2,720	.91	C
9	Corwin Springs---------	3,300	14,000	4.2	.82	200/ 3.1	65	.700	2,706	.88	C
10	Little Trail Creek-----	3,600	18,500	5.1	.84	300/ 5.6	54	.581	3,024	.82	U
11	Black Canyon-----------	3,550	[1]25,000	7	.88	500/ 6.8	74	.800	3,124	1.16	E
12	Little Cottonwood------	3,800	[1]32,000	8.4	.90	400/ 5.4	74	.804	3,420	1.27	E
13	Buffalo Creek----------	3,900	[1]42,000	11	.92	300/ 4.5	67	.723	3,588	1.20	E
14	Lamar R.S.-------------	3,900	[1]32,000	8.5	.90	500/15	33	.362	3,510	[2] .59	U
15	Amphitheater Creek-----	3,800	[1]15,000	4	.81	250/ 5.5	45	.493	3,078	[2] .71	U
16	NE. Entrance-----------	3,600	[1]10,000	2.7	.72	550/ 9	61	.663	2,592	[2] .79	U
17	Broadwater Lake--------	2,600	[1]30,000	12	.93	400/ 4.5	89	.964	2,420	[2]1.08	U
18	Slough Creek C.G.------	4,000	[1]40,000	10	.92	500/ 9	56	.603	3,680	[2]1.03	E
19	Slough Creek Ranch-----	4,150	[1]30,000	7	.88	500/10.4	48	.522	3,652	[2] .89	U
20	Buffalo Fork-----------	3,050	[1]25,000	8	.90	500/ 6.3	79	.861	2,745	1.10	E
21	Hellroaring Creek------	3,300	[1]30,000	10	.92	450/ 6	75	.814	3,036	[2]1.04	E
22	------do---------------	3,050	[1]18,000	6.0	.87	350/ 5.4	65	.703	2,650	.86	U
23	Mammoth----------------	3,300	21,000	6.4	.87	400/ 4.9	82	.886	2,871	1.18	E
24	Blacktail Plateau------	2,550	[1]40,000	16	.95	400/ 4.4	91	.986	2,422	1.11	E
25	Tower Cr. amphitheater-	1,900	[1]10,000	5.3	.85	800/10	80	.868	1,615	[2] .65	U
26	Lava Creek-------------	2,750	[1]25,000	9	.91	500/ 5.9	85	.920	2,503	1.07	E
27	Upper Lava Creek-------	1,850	[1]35,000	19	.95	500/ 4.8	104	1.130	1,758	.92	U
28	Swan Lake Flat---------	2,200	[1]30,000	14	.94	400/ 3.8	105	1.142	2,068	1.09	E
29	Winter Creek-----------	2,550	[1]35,000	14	.94	300/ 2.9	103	1.122	2,397	1.25	U
30	Gardners Hole----------	2,100	[1]35,000	17	.95	500/ 4.2	119	1.292	2,016	1.21	E
31	Fawn Creek-------------	1,800	[1]10,000	5.6	.86	500/ 2.8	179	1.937	1,548	1.39	E
32	Specimen Creek---------	320	1,500	4.7	.83	500/ 1.05	476	5.153	265	.63	C
33	Upper Specimen Creek---	900	3,000	3.5	.78	500/ 1.4	357	3.870	702	1.26	U?
34	------do---------------	750	8,000	10	.92	500/ 1.5	333	3.612	690	1.15	E
35	------do---------------	1,000	[1]10,000	10	.92	500/ 1.8	278	3.011	920	1.28	E
36	Fan Creek, overflow----	800	4,300	5.4	.85	500/ 1.7	294	3.188	680	1.01	U
37	Lower Gallatin---------	550	2,100	3.8	.80	500/ 1.3	385	4.166	440	.85	C
38	------do---------------	1,000	7,000	7	.88	500/ 2.1	238	2.582	880	1.05	U
39	------do---------------	1,500	6,000	4	.81	1,000/ 4.8	208	2.259	1,215	1.27	E
40	Grayling Creek---------	450	3,500	8	.90	500/ 1.5	333	3.612	405	.68	C
41	------do---------------	800	3,000	3.8	.80	500/ 1.4	357	3.870	640	1.15	C
42	Gneiss Creek-----------	500	2,100	4.2	.82	500/ 1.3	385	4.166	410	.79	C
43	Cougar Creek-----------	650	8,000	12	.93	1,000/ 3.4	294	3.188	605	.89	C
44	Tuff Creek-------------	800	[1]13,000	16	.95	1,000/ 3.4	294	3.188	768	1.14	U
45	Xmas Tree Park---------	950	[1]16,000	17	.95	500/ 2.2	227	2.464	903	1.03	U
46	Madison Canyon, mouth--	500	4,200	8.4	.90	500/ 1.3	384	4.166	450	.87	C
47	Mount Haynes-----------	1,000	2,400	2.4	.70	500/ 2.4	208	2.260	700	.73	C
48	Madison Junction-------	1,450	6,800	4.7	.83	500/ 3.4	147	1.595	1,204	.89	U
49	Gibbon River-----------	1,750	[1]15,000	9	.91	300/ 2.0	150	1.627	1,592	1.20	E
50	Canyon Creek-----------	2,050	[1]20,000	10	.92	400/ 3.0	133	1.446	1,886	1.26	U

[1]Half width based on projection of valley walls to ice surface.

[2]Minimum value for basal shear stress because ice thickness is a minimum value.

[3]C, compressing (decelerating) flow; U, uniform flow; E, extending (accelerating) flow. Queried where uncertain.

EQUILIBRIUM-LINE ALTITUDE

The ELA (equilibrium-line altitude), approximately the same as snowline, separates the accumulation and ablation areas of a glacier. There are several ways of estimating the full-glacial ELA. In the following discussion, the results of several methods are described and their relevance to the ELA of the northern Yellowstone glacier given; it is concluded that the glaciation limit provides the best estimate of the ELA for the Yellowstone icecap situation.

A method of estimating the ELA on mountain-valley glaciers is by assuming it to be the same as the altitude of the floors of the lowest cirques. In the areas immediately west of the northern Yellowstone icecaps, that altitude is 2,620 m (8,600 feet); this provides only a minimum value for the Yellowstone icecap because the ELA for valley glaciers is lower, generally 100–400 m (Østrem, 1966a, p. 137), than that appropriate for nearby icecaps; this is due to shading by the cirque and valley walls and greater accumulation due to wind drift of snow.

A rather crude method assumes that, along the sides of the glacier, the ELA lay below dominantly erosional glacial features and above depositional glacial features. In the Gardiner area the flanks of nunataks show only features of glacial erosion at altitudes above 2,680 m (8,800 ft), whereas a scattering of erratics are found up to that altitude.

Still another method assumes that the altitude of the lowest uplands that supported local icecaps is the altitude of full-glacial snowline. The lowest local icecaps in the northern Yellowstone area formed on the uplands in the headwaters of Specimen Creek between 2,710-2,860 m (8,900-9,400 ft ft) in altitude.

The AAR (accumulation-area ratio) method depends upon determining the area covered by a former glacier and assuming a specific AAR (Meier and Post 1962; Porter, 1975). Porter (1975) summarized data indicating that the equilibrium AAR's for *maritime* glaciers in western North America is between about 0.55 and 0.65 Meier and Post (1962, p. 70) cautioned that because piedmont glaciers and icecaps have very asymmetric area-to-altitude distributions, their AAR's may differ greatly from mountain-valley glaciers. The AAR method of estimating the ELA was not used because Yellowstone was glaciated by icecaps rather than by mountain-valley glaciers, and the area-altitude distribution of the northern Yellowstone glacier differed in having both a greater altitude range and detailed shape from the kinds of glaciers used to establish the relation between ELA and ARR.

I think that the ELA on *icecaps* such as the northern Yellowstone glacier is best approximated by the glaciation limit—the average altitude of the lowest nearby peaks that supported glaciers and the highest nearby peaks that did not. On the one hand, for valley glaciers, the ELA is commonly 100 to 400 m below the glaciation limit, largely due to shading and wind drift into cirques. On the other hand, on small mountain icecaps, the ELA may be slightly above the glaciation limit (Andrews, 1975, fig. 4-1C), because of wind deflation of snow. On large mountain icecaps, where the effects of shading and wind drifting are not important, the ELA should approach the altitude of the glaciation limit. If one assumes the ELA on icecaps is the same as the altitude of the climatic snowline, then the ELA may be about 100 m lower than the glaciation limit (Charlesworth, 1957, p. 11; Østrem, 1966, p. 129). Because the difference, if any, between the local glaciation limit and the ELA on adjacent icecaps is not well-established, I have assumed glaciation limit to be the same as ELA for purposes of mass-balance and accumulation-area-ratio calculations in northern Yellowstone.

The altitude of the glaciation limit in northern Yellowstone is as follows: northwest of Gardiner, Mont., and adjacent to the northern Yellowstone glacier, 2,830 m (9,300 ft); west and south of the Gallatin Range, 2,740–2,770 m (9,000–9,100 ft); and 15 km southwest of the map area, 2,620 m (8,600 ft). The glaciation limit shows a decrease toward the southwest, which is presently, and was probably in the Pleistocene, the direction of incoming winter storms, and also increasing snow depth on the Yellowstone plateau.

ACCUMULATION-AREA RATIO

The AAR (accumulation-area ratio) is the ratio of a glacier's accumulation area to its total area (Meier and Post, 1962, p. 70). The AAR for the northern Yellowstone glacier as portrayed on plate 1*A* is 0.75 (2,555 km^2/3,405 km^2; table 5). This AAR is also shown graphically on the left side of figure 49 by the ratio of the area above the ELA to the total area. This AAR of 0.75 is higher than the 0.6 ± 0.05 range of values that Porter (1975, p. 35) concluded was representative of mountain glaciers in New Zealand. This difference probably results both from the fact that the northern Yellowstone glacier was fed from icecaps and from the fact that it had a strongly skewed distribution of area with altitude, as shown on left side of figure 49. The effect of this skewed area-altitude distribution on the AAR is more completely evaluated by calculation of net balance in the next section.

Another way of appreciating this difference in range of AAR is to calculate the apparent ELA for the northern Yellowstone glacier, assuming an AAR of 0.6 ± 0.05 and given the area-altitude distribution from

plate 1*A* (table 5; fig. 49). This results in an apparent ELA of 3,010±50 m (9,875±150 ft) and places the ELA 180 m *above* the glaciation limit, which is not reasonable. A more fundamental way of examining the relations between the accumulation and ablation areas of the northern Yellowstone glacier is to calculate volumes, or mass balance, rather than areas (fig. 49).

The smaller glaciers in the northern Yellowstone area had more typical area altitude distributions, and their AAR's are as follows:

Gneiss Creek–Cougar Creek glacier	0.6
Gallatin River glacier	0.55–0.6
Fan Creek glacier	0.55–0.6
Specimen Creek glacier	0.6

MASS BALANCE

A fundamental parameter for evaluating the ELA, the AAR, and the glacial reconstruction itself (pl. 1*A*) is the estimated net balance on an annual basis of the northern Yellowstone glacier. The net balance of a glacial system can be estimated given the following: (1) the area-altitude distribution, (2) the ELA, and (3) the *specific* net balance as a function of altitude above or below the ELA. The area-altitude distribution can be derived from the ice-surface contours (pl. 1*A*); it is shown by the hypsometric bar graph on the left side of figure 49. The ELA on the surface of the icecap, as approximated by the glaciation limit, is 2,835 m (9,300 ft).

The point of this section on quantitative evaluation is to apply independent checks from modern glaciological studies to the field reconstruction of the glacial system in Pinedale time (pl. 1*A*). Because exact values for the specific net balance for the Yellowstone area cannot be inferred from studies of modern glaciers, I have used the following approach: the range of likely values are bracketed by estimates of the highest and lowest reasonable values of specific net balance, and an intermediate curve is drawn giving my best estimate of the specific net balance. These estimates were made from study of mass balance estimates in the literature, from consultation with colleagues, especially Mark Meier, and from examination of modern climatic data, especially snow-survey measurements. These estimates of specific net balance, when applied to the glacial reconstruction, represent a kind of sensitivity analysis that can be considered from two points of view. First, given the range of values, does the total net balance below the ELA cancel that above the ELA for the reconstructed northern Yellowstone glacier? Second, what were the actual volumes of total net balance above and below the ELA, and what do these mean in terms of ice and melt-water discharge?

SPECIFIC NET BALANCE ESTIMATES

Although the estimates are presented as three curves, they are discussed in six parts because the segments of each curve above and below the ELA require separate consideration. It is not intended that, for example, the low estimate for the ablation area *must* connect with the same curve for the accumulation area.

The "high" estimate (fig. 49) is a generalized curve by S. C. Porter (written commun., 1976), and extended by me. It is drawn by visually averaging the specific net balance curves of nine maritime glaciers in the Pacific Northwest; many of these curves of specific net balance used for this generalization are shown in Meier and others (1971, fig. 1). Although appropriate for evaluating maritime glacial reconstructions, the net balance gradient below the ELA of −125 cm/100 m probably represents a maximal value for the continental climate of Yellowstone.

For the high estimate of specific net balance above the ELA (fig. 49), a maximum value of +2 m seems high but not impossible for the Yellowstone area. Modern snow survey measurements provide some insights into the moisture regimen of this area. In the southwest corner of Yellowstone Park, on the Pitchstone Plateau, at an altitude of only 2,630 m (8,640 ft), the snowpack on April 1 contains about 170 cm water. For the western Beartooth uplift and the western Absaroka Mountains of the Yellowstone area, the April 1 snowpack at altitudes of about 2,800 m (9,100–9,400 ft), the highest stations observed, contains about 100 cm of water (Peak and Clagett, 1973, p. 67, 135, and 186). For the northern Yellowstone area, projection of altitude-moisture trends to 3,280 m (10,750 ft) suggests 170-cm water content in the snowpack—the same as that derived from the "high" estimate of specific net balance (fig. 49; table 5). For a given altitude, the snowpack in the northern Yellowstone area has about three times as much water as the snowpacks in the Bighorn Mountains and the Wind River Range. Thus, although far inland, the Yellowstone uplands are (and probably were) uncommonly wet, apparently because the Snake River Plain serves as a low-altitude conduit permitting moisture-laden storms to reach the Yellowstone area.

The "low" estimate of specific net balance below the ELA describes a net balance gradient of −70 cm/100 m. For glaciers whose ablation is not reduced by orographic shading, northerly slope, debris mantling or occurrence at high latitudes, net balance gradients below the ELA lower than −70 cm/100 m are uncommon, with gradients averaging about 90 cm/100 m (Meier and others, 1971, fig. 1A; Østrem and Liestol, 1964; Pytte and Ostrem, 1965; Østrem, 1966b; Mercanton,

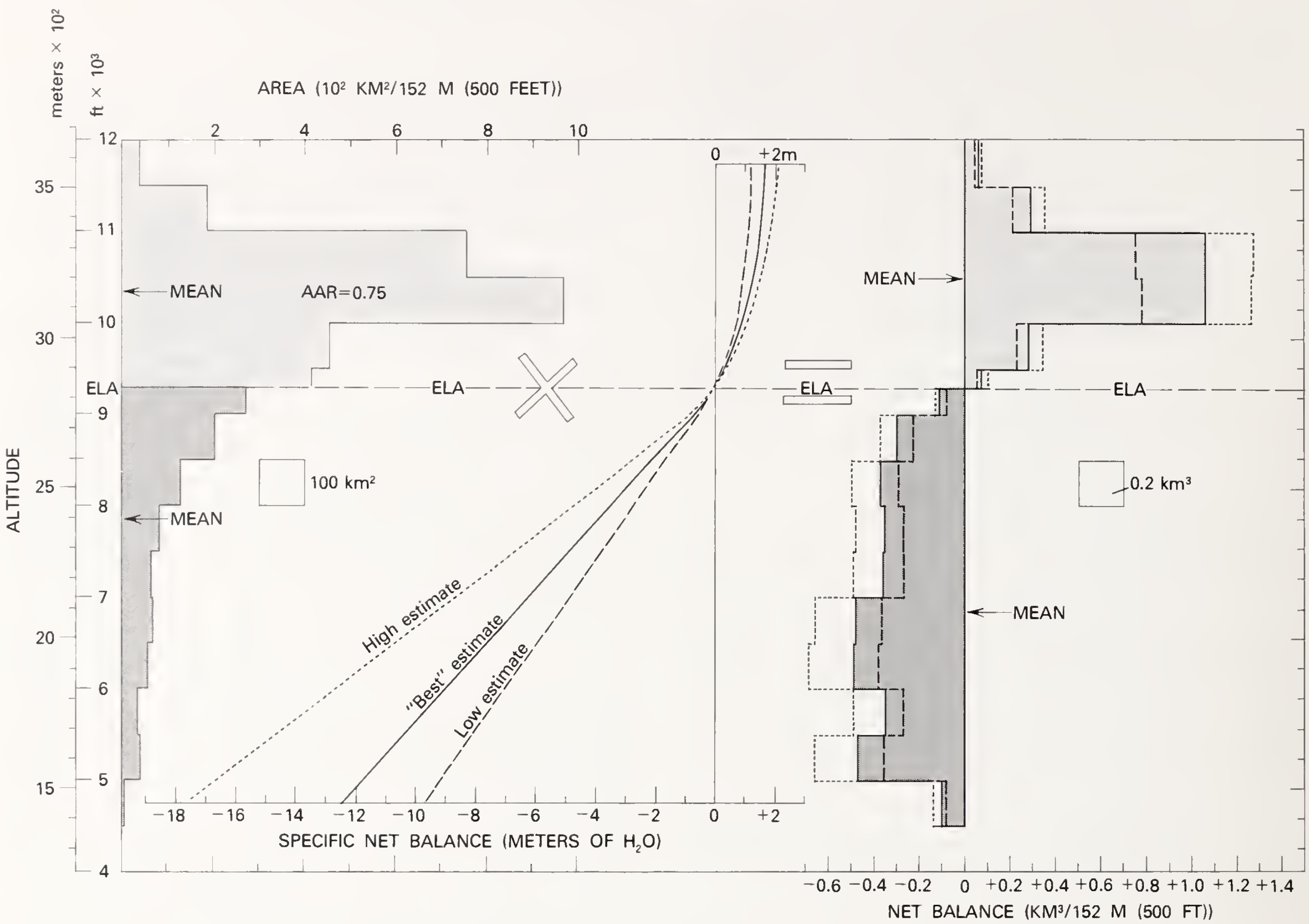


FIGURE 49.—Mass balance of the northern Yellowstone glacier, which terminated in the Yellowstone valley and headed at ice divides on icecaps in and near the park. On the left is a hypsometric bar graph showing the area-altitude distribution of the surface of the icemass defined on plate 1*A*. The ELA (equilibrium-line altitude) is inferred to be the same as the glaciation limit, 2,835 m (9,300 ft). The AAR (accumulation-area ratio) of 0.75 is derived from the areas of the accumulation area and total area given in table 5. The high and low estimates of specific net balance indicate the probable range of Pleistocene values. On the right side of the figure, the net balance is the product of the area of each altitude increment multiplied by three estimates of specific net balance. For example, for the altitude interval 3,353–3,200 m, the area (746.5 km²) multiplied by specific net balance (+1.0 m) equals the total net balance (0.75 km³) (table 5). Net balance is in terms of water equivalent.

1916, p. 87; Schul'tz, 1962, p. 180; Meier and Post, 1962). Net balance gradients less than −70 cm/100 m have been observed (Schytt, 1967; Meier and others, 1971; Pytte and Østrem, 1965) but these low values are for glaciers where I infer that total ablation has been reduced by orographic shading, insulation by debris mantle, northerly slope, or location at high latitude.

The "low" estimate of specific net balance above the ELA approaches a high value of +1.2 m (fig. 49). Nearly all sizeable glaciers in near-equilibrium conditions have maximum accumulation values of at least +1 m (Meier and others, 1971, fig. 1; Østrem and Liestol, 1964; Pytte and Østrem, 1965, fig. 28; Østrem, 1966b). Modern snowcourse studies suggest that present the northern Yellowstone area is easily capable of providing the precipitation required by the "low" estimate.

I based the "best" estimate for the specific net balance below the ELA on analogy with the Saskatchewan Glacier of Alberta and the Fedchenko Glacier of the Pamirs of central Asia, because these are the most analogous and the best studied large continental glaciers at temperature latitudes. These glaciers have net balance gradients below the ELA slightly greater than −90 cm/100 m over total altitude intervals of 800 m and 1,500 m respectively (Meier and others, 1971, fig. 1; Meier, 1960, fig. 7; Schul'tz, 1962, p. 180; Budd and Jenssen, 1975, fig. 7).

Other glaciers approximate the Yellowstone situa-

TABLE 5.—*Values used in calculations of mass balance*

Altitude interval			Low estimate		Best estimate		High estimate	
Meters	Thousand feet	Area (km²)	Specific net balance (m)	Volume (km³)	Specific net balance (m)	Volume (km³)	Specific net balance (m)	Volume (km³)
3,658-3,505	12.0-11.5	33.70	1.2	+0.04	+1.70	+0.06	+2.10	+0.07
3,505-3,353	11.5-11.0	185.29	1.15	.21	1.55	.29	1.90	.35
3,353-3,200	11.0-10.5	746.51	1.0	.75	1.42	1.06	1.70	1.27
3,200-3,048	10.5-10.0	968.50	.8	.78	1.10	1.06	1.30	1.26
3,048-2,896	10.0-9.5	454.78	.5	.23	.62	.28	.75	.34
2,896-2,835	9.5-9.3	166.02	.15	.02	.25	.03	.25	.04
Total above ELA--		2,554.8		+2.03		+2.78		+3.33
2,835-2,743	9.3-9.0	163.01	-0.31	-0.05	-0.41	-0.07	-0.5	-0.08
2,743-2,591	9.0-8.5	200.31	-1.17	-.23	-1.51	-.30	-1.85	-.37
2,591-2,438	8.5-8.0	128.72	-2.24	-.29	-2.88	-.37	-3.85	-.50
2,438-2,286	8.0-7.5	82.41	-3.31	-.27	-4.25	-.35	-5.80	-.48
2,286-2,134	7.5-7.0	63.37	-4.37	-.28	-5.62	-.36	-7.75	-.49
2,134-1,981	7.0-6.5	68.58	-5.44	-.37	-7.00	-.48	-9.70	-.66
1,981-1,829	6.5-6.0	58.95	-6.51	-.38	-8.37	-.49	-11.70	-.69
1,829-1,676	6.0-5.5	36.10	-7.57	-.27	-9.74	-.35	-13.65	-.49
1,676-1,524	5.5-5.0	42.11	-8.64	-.36	-11.11	-.47	-15.65	-.66
1,524-1,372	5.0-4.5	7.82	-9.71	-.08	-12.48	-.10	-17.70	-.14
Total below ELA--		851.38		-2.58		-3.34		-4.56
Total area-------		3,406.18						

tion to a lesser degree because they differ from the northern Yellowstone glacier in size, orographic situation, and latitude. The Gulkana Glacier of central Alaska has a net balance gradient of −90 cm/100 m over an altitude interval of 350 m in its ablation area upglacier from its debris-mantled snout (Meier and others, 1971, fig. 1). Where not highly shaded, the Peyto Glacier of Alberta has a net balance gradient of −90 cm/100 m in its ablation area (Østrem, 1966b, fig. 14).

Values for the "best" estimate of net balance above the ELA are more difficult to infer, because of the paucity of data. Where data do exist, wind drift and shading by headwalls complicate comparison with the open, nearly horizontal upper surface of the vast northern Yellowstone glacier. For the open, nearly horizontal upper basins of the Gulkana Glacier of interior Alaska in 1966 and 1967, the net accumulation averaged a maximum of about +1.4 m at an altitude of 500 m above the ELA (Meier and others, 1971, pl. 5; Tangborn and others, 1977, pl. 4*E*). In interior Canada, net accumulation on the Peyto Glacier is +2 m at heights 250–450 m above the ELA, and on the Woolsey Glacier it is +1.0 m and +1.3 m at heights of 70 and 200 m, respectively, above the ELA (Østrem, 1966b). For both the Peyto and Woolsey Glaciers, wind drift is of some importance; and the glaciers are more northerly and the Woolsey is less continental than the Yellowstone. The source area of the Saskatchewan Glacier in the Columbia Icefields is of similar form to the Yellowstone icecaps. As much as +3 m net accumulation is shown for this area (Meier and others, 1971, fig. 1), but this amount appears anomalous compared to more maritime areas; and the value is based only on one snow pit at one time and may not be representative (M. F. Meier, written commun., 1976). In summary, the basis for the "best" estimate of the specific net balance in the accumulation area of northern Yellowstone was the Gulkana Glacier, the Peyto and Woolsey Glaciers (minus a little to account for wind drift), and a median between the "high" and "low" estimates (fig. 49). This curve is not in excess of the moisture predicted by plots of present spring snowpack data against altitude discussed earlier (p.73). M. F. Meier (oral commun., 1976) considers a likely value for *average* accumulation to be about +1 m; table 6 shows that the "best" estimate yields an average of +1.09 m.

TABLE 6.—*Summary of results of the mass balance calculations*

Mass balance	Balance estimates		
	Low	Best	High
Total net balance above the ELA	$+2.0$ km³	$+2.8$ km³	$+3.3$ km³
Total net balance below the ELA	-2.6 km³	-3.3 km³	-4.5 km³
Apparent net balance	-0.6 km³	-0.5 km³ $-(0.0$ km³)[1]	-1.2 km³
Specific net balance			
Mean of values in accumulation area.	$+0.79$ m	$+1.09$ m	$+1.30$ m
Mean of values in ablation area .	-3.03 m	-3.92 m	-5.36 m

[1]If the ELA is lowered about 50 m, accumulation and ablation are equal.

MASS BALANCE RESULTS

The high, low, and best estimates of specific net balance, multiplied by the area-altitude distribution, yield the volume of annual accumulation and ablation for each altitude interval. These values are shown by the dashed, solid, and dotted lines on the right side of figure 49, and given in tables 5 and 6. Under equilibrium conditions at the Pinedale glacial maximum, the total net balance above the ELA should equal the total balance below the ELA. As shown by table 6, the estimates of total net balance above the ELA range from $+2.0$ to $+3.3$ km³ and the estimates of total net balance below the ELA range from -2.6 to -4.5 km³. The "best" estimate yields a total net balance above the ELA of $+2.8$ km³ and a total net balance below the ELA of -3.3 km³—a difference of -0.5 km³, or 15 percent. The reconstructed northern Yellowstone glacier can be balanced well within the tolerances of the calculations. For example, within the limits of the estimates, each may have been as high as 3.3 km³ or as low as 2.6 km³, a range of 25 percent, and still have balanced each other. Based on the "best" estimate, total net balance above and below the ELA equal to about 3 km³ is the most reasonable.

The most significant uncertainty in the mass balance calculation is the ELA estimate. If the ELA were lowered by only about 50 m, then accumulation would exactly balance ablation in the case of the "best" estimate of specific net balance. The ELA estimate is based on the glaciation limit, which itself can be determined to only about ±50 m; in addition some think the ELA for icecaps is 100 m lower than the glaciation limit. Thus, by changing the ELA well within permissible limits, an exact balance between accumulation and ablation can be obtained.

The results of the mass balance calculation also shows that the apparently high AAR (0.75) poses no problem. In fact, if the ELA is placed 50 m lower in order to achieve an exact balance between accumulation and ablation, then the AAR is increased to 0.78.

This calculation of mass balance does not provide as critical a test of the glacial reconstruction as does basal shear stress, but it does show that the reconstruction is reasonable. If the area-altitude distribution above or below the ELA (fig. 49; pl. 1*A*) were changed by 25 percent, or if the ELA were changed by about 200 m, then the calculations would not produce a reasonable balance.

GLACIER ICE DISCHARGE

As outlined by models A, B, and C in table 7, three different approaches may be used to estimate the discharge and velocities of former glaciers. A reach was selected for these calculations to see how discharges based on different lines of reasoning compared. The best reach of the northern Yellowstone outlet glacier for discharge calculations is the rather uniform reach centered about 5 km upstream from Yankee Jim Canyon, here called the Cedar Creek transect.

For flow *within* a glacier, a velocity distribution shown diagrammatically by Nye (1965, fig. 8a) was used to estimate the relation between centerline velocity and average velocity. For basal sliding, the centerline velocity due to basal sliding is the *same* as the average velocity along the centerline of the glacier surface. But where the ice thins towards the sides of a glacier, the amount of basal sliding apparently diminishes considerably. This decrease produces strong shear stress gradients and the crevasses common on glaciers whose movement is dominated by basal sliding (fast glaciers and surging glaciers). This

TABLE 7.—*Glaciological values for flow at the Cedar Creek transect*

[Reconstructed glacial parameters at the Cedar Creek transect: Ice surface altitude, 2,500 m (8,200 ft); bedrock altitude at center line, 1,525 m (5,000 ft); thickness, 975 m; width, 9.7 km; cross-sectional area, 5 km²; ice-surface slope, 0.72°]

	Model A[1] (based on analogy to modern, smaller glaciers)	Model B[2] (based on flow-law and basal sliding calculations)	Model C (based on discharge required by net ablation down-glacier)
Centerline velocity (km per year)			
Resulting from flowage within ice--	---	[3]0.065(0.190)	[5]0.065(0.190)
Resulting from basal sliding-------	---	[3] .076?	[6] [7] .63
Total---------------------------	0.2	[3] .14 (0.27)	.69 (0.82)
Average velocity (km per year)			
Resulting from flowage within ice--	---	[4]0.04(0.12)	[4] [5]0.04(0.12)
Resulting from basal sliding-------	---	[7] .05?	[6] .42
Total---------------------------	0.1-0.15	.09(0.17)	.46(0.54)
Glacial discharge (km³ per year)			
Resulting from flowage within ice--	---	0.2(0.6)	[5]0.2(0.6)
Resulting from basal sliding-------	---	.25?	[6]2.1
Total---------------------------	0.5-0.7	.45(0.85)	[8]2.7

[1]Only totals given; other data generally not available.
[2]Queried entries indicate great uncertainty in basal sliding calculations.
[3]See text for calculations. Value in parentheses is maximumum limit.
[4]Estimated from velocity distribution diagrams of Nye (1965, fig. 8a).
[5]Flow-law calculation is assumed to approximate value.
[6]These values calculated from difference between discharge based on down-glacier net ablation and maximum value discharge estimated to result from flowage within the ice.
[7]For basal sliding, the average velocity is estimated by me to be about two-thirds the sliding velocity along the centerline.
[8]Net ablation down glacier, from table 5.

basal-sliding-velocity gradient was incorporated in my estimates of the contribution of basal sliding to discharge (table 7).

Model A is based on analogy to modern glaciers. For three well-studied modern glaciers with nearly horizontal basal surfaces in their ablation area, the ratio of annual centerline velocity to depth is smaller than 1:4. Actual values are 1:6–7.5 for the Saskatchewan Glacier (Meier, 1960); 1:4.5–9 for the Athabasca Glacier (Paterson, 1970); and 1:5–8 for the Fedchenko Glacier (as reported in Budd and Jenssen, 1975, fig. 7). An extrapolation of this data predicts a centerline velocity no greater than 215 m (1/4.5×975 m) per year at the Cedar Creek transect. As shown in table 7, this would result in average velocities of about 0.1 to 0.15 km per year and glacial discharges through the cross-sectional area (5 km^2) of about 0.5 to 0.7 km^3 per year.

Model B uses flow law and basal sliding relations derived from laboratory and field study of glaciers.[5] Hodge (1972) evaluated flow-law parameters determined from studies of a number of different temperate glaciers. In most cases a simple power-low relation holds:

$$\dot{v}=A\tau^n$$

where $\dot{v}$ is the effective shear strain rate and A and n are the flow-law parameters (Nye, 1953). According to Hodge's analysis, the original estimates of Nye (1953) that A is 0.173 and n is 3.07 are close to the mean values of 17 different experiments or analyses, and the estimates of Shreve and Sharp (1970), that A is 0.55 and n is 3.3, represent the least viscous cases (producing most rapid flow). The speed of glacial flow due to internal deformation, u_d, is given by integrating $\dot{v}$ from the surface to the bed of the glacier, thus

$$u_d=2Ah\tau^n(n+1)^{-1}.$$

For the Cedar Creek transect, the flow due to internal deformation is 65 m per year, using the parameters of Nye (1953); the most rapid value is 189 m per year using the parameters of Shreve and Sharp (1970). Another analysis using data from many glaciers related τ and u_d/h according to an empirical relation (Budd and Jenssen, 1975, p. 271):

$$u_d/h=0.0856\tau^{2.5}.$$

For the Cedar Creek transect, u_d=68 m per year.

It is difficult to provide a good estimate of basal sliding. Paterson (1970) and Hodge (1972) have shown that existing theories do not seem to match observed values. Paterson (1970, p. 62) showed a possible empirical relation between sliding velocity and ice thickness, but states that "such a relation cannot be true everywhere." His relation, extended beyond the field of his data points, yields a sliding velocity at the Cedar Creek transect of 76 m per year, but the resulting value cannot be considered reliable.

Thus for model B (table 7), the centerline velocity (140 m, not exceeding 270 m, per year), and discharge (0.45, not exceeding 0.85, km^3 per year) are clearly less than 1 km^3 per year.

Model C is based on the volume of ice required to flow across the Cedar Creek transect to replenish the ice lost downglacier by net ablation. The "best" estimate indicates this amount was about 2.7 km^3 per year (total net balance below 2,500 m, table 5, fig. 49). The net ablation is affected by (1) the downglacier area-altitude distribution, (2) the ELA, and (3) the curve of specific net balance. Use of my estimates for the *minimum* likely value for each of the above yields net ablation downglacier of 1.9 km^3 per year, a reduction of 30 percent. The flow-law calculations, even using the "most rapid" observed parameters, show that discharge due to flowage within the glacier is only a small fraction of this. In order to meet the discharge required by downglacier ablation, basal sliding in the order of half a kilometer per year is required (table 7). Thus, if the mass balance values are accurate within a factor of 2, then basal sliding exceeding the velocity due to flow within the glacier is required (model C, table 7).

In order to better appreciate whether basal sliding may have been dominant, the following estimate of water balances is made from modern streamflow measurements to see if these independent data support the mass balance conclusion that ice discharges at the Cedar Creek transect of several cubic kilometers per year are reasonable. The streamflow from that part of the present Yellowstone River drainage that was occupied by the northern Yellowstone glacier is closely approximated by the difference between the annual streamflow measured just upstream from the Cedar Creek transect at Corwin Springs and that at the outlet from Yellowstone Lake (U.S. Geological Survey, 1974, p. 427, 436). This value is 1.58 km^3 per year, and it represents an average runoff of 38 cm from the area of the watershed. The present evaporation and transpiration losses from this kind of terrain are about an additional 38 cm per year (Leaf, 1975, p. 3), indicating average precipitation of about 76 cm per year. During full-glacial time the ice surface averaged about 600 m higher than the present land surface. A common equation for the increase in precipitation with altitude is about 6 cm per 100 m (Landsberg, 1958, p. 176), or an additional 36 cm, for this increase of 600 m. If one assumes that values of precipitation during glacial times were similar for a given altitude in northern Yellowstone to those at present, then the estimated average annual precipitation was about 112 cm. This

[5] The calculations and text for model B are by Mark Meier (written commun., 1977).

assumption is not unreasonable, for pluvial lakes were widespread in the western United States during glacial times, and water balance calculations for these lake basins indicate that precipitation was not significantly reduced (G. I. Smith, oral commun., 1977). From about 6 to 9 cm of water is lost by evaporation from present Rocky Mountain seasonal snowpacks (Leaf, 1975, p. 16); assuming annual losses from the Pleistocene icecap were between two and three times this amount, about 20 cm annual loss would result. Thus, based on these estimates, the average precipitation minus the loss by evaporation equals 92 cm or 2.4 times the modern runoff. This would yield a total discharge of ice and water at the Cedar Creek transect of 3.8 km^3 per year. This estimate compares favorably with the 3 km^3 per year estimate of glacial discharge beneath the ELA (table 5; fig. 49), and it supports the conclusion that annual discharge was of the magnitude of several cubic kilometers and not just a fraction of a cubic kilometer, as required by the velocity calculations of models A and B, table 6.

BASAL SLIDING

A persistent problem in glacial geology and glaciology is the amount of basal sliding of glaciers. Flow within the glacier is relatively well defined by the flow laws of ice, but quantification of basal sliding remains elusive, as is dramatically shown by the problem of surging glaciers. With regard to the Yellowstone reconstruction, the fact that the discharge of the outlet glacier was about four times larger than that predicted by velocity calculations may not represent a serious problem. Put simply, the problem is whether the 75 m/yr estimate of basal sliding (table 7, model B) accurately represents the Yellowstone situation, or whether basal sliding was actually as much as 600 m a year (model C).

Some glaciers smaller than the Yellowstone outlet glacier with rather normal surface slope–thickness relations may surge periodically with basal sliding rates of several kilometers per year for short periods, more than an order of magnitude faster than the rate required for the northern Yellowstone glacier by the mass balance estimates. Some outlet glaciers in Greenland flow continuously at speeds of several kilometers per year; it is presumed that most of this speed is due to basal sliding. Thus, the crux of the problem of quantitatively evaluating the northern Yellowstone glacial discharge rests with the uncertainties of determining basal sliding.

W. F. Budd (1975, p. 3, fig. 2) concluded that three classes of glaciers exist because two different modes of flow exist:

> "Ordinary" glaciers do not have sufficient mass flux, for the given bedrock profile, to go beyond the "slow mode" in which the basal stress and velocity increase together as the glacier builds up to steady state.
>
> "Fast" glaciers have sufficient flux to remain continuously in the "fast mode" with high velocities and relatively low basal stress.
>
> "Surging" glaciers have sufficient flux rate to reach the fast mode but not sufficient to maintain it, and thus develop a periodically oscillating state between the fast and slow modes with gradual buildup and rapid drainage.

For fast glaciers, Budd (1975, fig. 2) plots 21 centerline velocity measurements that range from 300 to 4,000 m a year, and average about 1,500 m a year. The boundary between fast- and slow-mode glaciers lies between 150 and 400 m per year. Only for glaciers in the slow mode does he find that velocity is systematically related to thickness and slope. The thickness and slope of the reconstructed northern Yellowstone glacier place it either in the very upper limit of the field of slow glaciers or in the field of fast glaciers. The flux rate (centerline thickness×velocity) of the reconstructed Yellowstone glacier, 670×10^3m^2/yr (model C), is clearly in Budd's field for fast glaciers.

This exercise in checking the discharge computed from downvalley ablation against flow rates determined by ice thickness, slope, and cross section suggests that basal sliding dominated the total flow. Both a sensitivity analysis of the melting in the downstream ablation area and the discharge calculated from modern records of streamflow, evaporation, and change in precipitation with altitude indicate flows of at least 1.9 km^3. Thus the reach of the reconstructed northern Yellowstone outlet glacier between Gardiner and Yankee Jim Canyon is a good candidate for a "fast" glacier in the sense of Budd (1975). The scouring of bedrock to form a closed basin upstream from Yankee Jim Canyon requires considerable basal sliding. This basin would also restrict escape of subglacial water and thereby facilitate basal sliding. The hot springs between Gardiner and Corwin Springs would engender abundant subglacial water and thereby also favor basal sliding.

ESTIMATE OF MELT-WATER VOLUMES

It is commonly thought that rapid recession from a glacial maximum produces large quantities of melt water and concomitant outwash deposits. The calculation of ice discharge provides a method of evaluating the magnitude of increase in melt-water discharge accompanying rapid glacial recession. The effect of melt-water discharge on outwash deposition is more difficult to evaluate, for it depends in large part on the ratio of sediment supply to melt water; this discussion

does not address this ratio but only the relative volumes of melt water. The annual *net* balance below the ELA indicated by the "best" estimate of specific net balance is 3 km^3 for the northern Yellowstone glacier (table 6). Based on comparisons with modern drainages containing glaciers, the total annual runoff from the northern Yellowstone glacier including melting in the accumulation area, precipitation on the ablation area, and nonglacial runoff from the basin, was probably at least twice the *net* balance below the ELA. For the Gulkana Glacier, total runoff was four times the net balance below the ELA (Tangborn and others, 1977, p. B18). If runoff is at least twice the net balance below the ELA, then the Yellowstone River discharge was at least 6 km^3. The total volume of the glacial system of the northern Yellowstone outlet glacier is approximated by multiplying the area (3,400 km^2) by the estimated average thickness (0.6 km). This yields a total of about 2,000 km^3. Glacial recession rapid enough to melt *half* this glacial system in only 2,000 years would add an additional 0.5 km^3 annually to the runoff. This addition is less than 9 percent of the estimated annual runoff during the glacial maximum. Thus, rapid meltback of the northern Yellowstone outlet glacier from the terminus to probably well beyond the Deckard Flats position near Gardiner (pl. 2) would produce only a small increase in runoff. Put another way, the normal full-glacial runoff dwarfs any addition to runoff due to rapid glacial recession. Most of the outwash of the northern Yellowstone outlet glacier is graded to the Eightmile and Chico end moraines (p. F29–F33; fig. 16); this also supports the conclusion that most of the outwash deposition occurred at or near full-glacial time, rather than during glacial recession.

SUMMARY OF QUANTITATIVE EVALUATION

Given a reconstruction of a glacial system based on field studies, calculations of basal shear stress and net budget provide independent checks of such a reconstruction, demonstrating whether a reconstruction is or is not physically reasonable.

Calculated values for primary basal shear stress of the glacial reconstruction shown on plate 1*A* range between 0.6 and 1.5 bars (fig. 49; table 4). These values are within the general limits observed on modern glaciers. The variation of the calculated values for the Yellowstone icemass is largely related to extending, uniform, or compressing flow.

Basal shear stress can be used to evaluate different glacial reconstructions of the same area. The lower of two alternative limits shown for Pinedale ice (U.S. Geological Survey, 1972a) yields values of primary basal shear stress between 0.0 and 0.15 bars. This is well below the normal range of 0.5–1.5 bars observed on modern glaciers, thus indicating that the lower alternate is not reasonable.

The accumulation area ratio (AAR) for the glacial system contributing to the northern Yellowstone outlet glacier (pl. 1*A*) is 0.75, about 0.1 higher than observed for mountain-valley glaciers. Because the Yellowstone system was an icecap and had a different area-altitude distribution than most mountain-valley glaciers, a calculation of net balance was made, partly to determine if the high AAR could be explained as a result of this area-altitude distribution.

The net balance is a function of the area-altitude distribution, the equilibrium-line altitude (assumed for large icecaps to be the same as the glaciation limit), and the specific net balance. A curve of specific net balance was constructed using curves of modern continental glaciers (fig. 49). The net balance determined by these parameters yields a glacial system for which net ablation is approximately equal to net accumulation—the condition necessary during a stable glacial maximum. These net-balance calculations indicate the plausibility of the reconstruction shown on plate 1*A* and suggest that the high AAR results from the area-altitude distribution of the northern Yellowstone glacier.

The net balance calculations were used to estimate the glacial discharge through the Cedar Creek transect of the northern Yellowstone outlet glacier. This discharge is several times that calculated for "normal" glaciers from flow-law formulas (table 7). Because the value of downvalley ablation is probably accurate within 30 percent, the alternative is that basal sliding dominated the flow of the Yellowstone outlet glacier.

Another ramification of the net balance and discharge calculations is that the relative increase in melt-water volume that would accompany rapid glacial recession would increase the estimated melt-water volume by less than 9 percent.

MODEL OF SPATIAL AND TEMPORAL RELATIONS BETWEEN SOURCE AREAS OF THE NORTHERN YELLOWSTONE GLACIER

The northern Yellowstone area offers an unusual opportunity to study the interrelations of glacial source areas of different character through time. This section attempts to model, by graphical and relative numerical means, the response of the different sources of the northern Yellowstone glacier through a glacial cycle. Most of the field information relates to full-glacial and deglacial times; information on the glacial buildup and oscillations near full-glacial times were generally

obliterated by conditions during the last full-glacial stand. However, glacial buildup was somewhat analogous, in reverse, to deglaciation.

Table 8 attempts to show in a semiquantitative fashion the relations between source areas of the northern Yellowstone glacier. Figure 50 shows the relative sizes and spatial relations of the different source areas. Column A of table 8 includes the area within each source region that is in the *present* drainage of the Yellowstone River whether or not it contributed to the northern Yellowstone glacier in full-glacial times. The values in column B are measured directly from plate 1*A*. The percent each source area makes up of the total source area of the northern Yellowstone glacier is also given in column B. The icecap on the Beartooth uplift formed just over half the source area of the northern Yellowstone glacier, whereas the icecaps on the Yellowstone plateau (II, III) together formed about 30 percent.

In column C, the ratio of B to A indicates the number of times larger the full-glacial source area was than the area of the present terrain above the full-glacial ELA. As can be seen from column C, the icecaps on the plateau (II, III) somehow expanded to large size, from terrain with little or no altitude estimated above the long-term ELA. Glacial buildup on the Yellowstone plateau was aided by inflow of ice from nearby ranges, thereby increasing the surface altitude. Also, the ELA may have dropped for short intervals significantly below the estimated full-glacial altitude of 2,835 m, and ice accumulated on higher parts of the plateau surface itself.

The particular spatial relations between the source areas of the northern Yellowstone glacier fostered glacial buildup and consequent increase in the area above the ELA. The ice divide at the head of the northern Yellowstone glacier outlines a rectangle that is closed except where the outlet glacier exits through its northwest corner. This almost centripetal flow pattern resulted in a general congestion that inhibited outflow of ice to the outlet glacier and promoted the buildup of ice surface to higher levels, thus adding to the glacial area above the ELA.

The orographic-snowfall effect (table 8, column D) is included to show the degree to which glacial buildup was self amplifying. The icecaps on the plateau (II, III) had strongest self-amplifying effect because they lay in the path of moisture-laden airmasses rising up onto the Yellowstone plateaus from the Snake River Plain. As ice built up on the plateau the resultant increase in altitude caused increasingly greater amounts of snowfall. This orographic-snowfall effect applies mainly to the time of glacial buildup.

Columns E and F of table 8 give my estimate of the

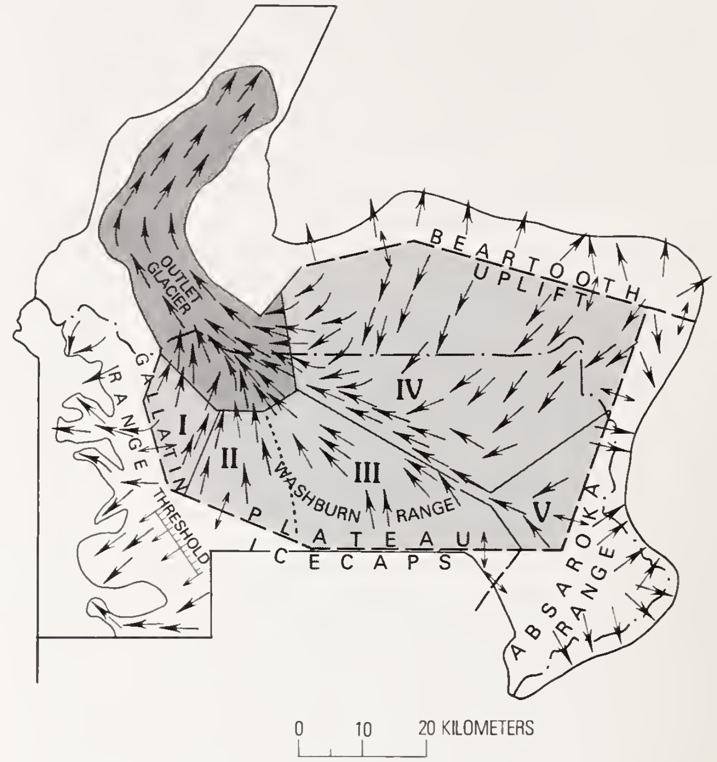


FIGURE 50.—The northern Yellowstone glacier, showing its major source areas and its outlet glacier. Heavy dashed line with double-headed arrows indicates ice divides. Smaller arrows, direction of ice flow. Boundary between areas II and III is dotted to indicate these two areas are both part of the plateau icecap.

relative productivity per unit area of the various source areas under full-glacial and waning glacial conditions. The differences in the magnitude of the numbers is intended to reflect the relative annual thickness of ice accumulation averaged over the full-glacial source areas (column B). Under full-glacial conditions the highest value is arbitrarily set at 10; under waning glacial conditions at 5. Waning glacial conditions refers to the situation in the recession about two-third of the way from the full-glacial to the Deckard Flats ice position. These numbers and the numbers derived from them (columns G, H, I, and J) are of course not quantitatively precise and are given only to aid the reader in understanding my conception of the relation between source areas, especially the changes from full-glacial to waning glacial times. For full-glacial conditions, two factors are most important to relative difference in average accumulation (column E): the altitude of the source area above the ELA (topographic contours, pl. 1*A*), and the position of the source area relative to source of moisture. But, under waning glacial conditions the altitude of the terrain becomes increasingly important to the sustaining of glaciers. The most important factor is the size of the area above the ELA relative to the size of the full-glacial source area (column B).

TABLE 8.—*Numerical model of the spatial and temporal relations among the source areas of the northern Yellowstone glacier*

[Values in parentheses in columns B, H, and I are the contribution of each source area, in percent, to the total for that column. For relative ice productivity, size of source areas same as in column B; relative thickness of accumulation on arbitrary scale of 1 (low) to 10 (highest full glacial conditions). Subjective estimates in columns E and F based on my ranking of the relative differences between source areas. In addition to the relative productivity of the different source areas indicated by mapping (pls. 1, 3, and 4), these estimates are also based on the average altitude of the different source areas and the location of the source areas relative to sources of precipitation]

	Buildup to full-glacial conditions				Relative ice productivity under full-glacial and waning glacial climate					
					Relative thickness of accumulation, entire source area			Relative indices of ice output		
	A	B	C	D	E	F	G	H	I	J
Source areas of the northern Yellowstone glacier	Area in km² of present terrain above 2,835 m (9,300 ft)	Area in km² of icecap surface above ELA in full-glacial time	B/A=Index to glacial buildup	Orographic-snowfall effect[1]	Full-glacial conditions	Waning glacial conditions	Sustaining factor F/E	B x E = Index of full glacial productivity	B x F = Index of waning glacial productivity	Incursion factor percent I/ percent H
I. Gallatin Range icecap--------- (south of Electric Peak).	35	120 (4)	4	Self-amplifying.	10	5	0.5	1,200 (5)	600 (6)	1.2
II. Plateau icecap between the---- Gallatin and Washburn Ranges.	0	235 (9)	∞	Strongly self-amplifying.	9	2	.2	2,115 (9)	470 (5)	.6
III. Plateau icecap flowing across the Washburn Range-Specimen Ridge area.	15	600 (22)	40	Strongly self-amplifying.	9	2	.2	5,400 (24)	1,200 (12)	.5
IV. Beartooth uplift icecap-------	230	1,425 (52)	6	Self-amplifying.	8	5	.6	11,400 (51)	7,125 (73)	1.4
V. Ice in the upper Lamar-------- drainage basin.	120	180 (7)	2	In snow shadow of plateau icecaps.	7	2	.3	1,260 (6)	360 (4)	.7
Local glaciers tributary to the-- northern Yellowstone outlet glacier.	50	150 (6)	3	Little effect	8	0.5	.05	1,200 (5)	75 (1)	.2
Totals-----------	450	2,710 (100)						22,575 (100)	9,830 (100)	

[1]Based on position relative to moisture sources and my assessment of the way the area acted during a glacial cycle.

The sustaining factor (column G), the ratio of the values in column F to those in column E, provides a measure of the relative ability of the different source areas to sustain glaciation under a waning glacial climate. The Beartooth uplift ranks highest because it contains the most high terrain, and also because the snowshadow effect of the plateau icecaps on the Beartooth uplift would have diminished as the plateau icecaps downwasted. The plateau icecaps have low values because they were located mostly on terrain with altitudes below 2,500 m, well below the waning glacial ELA, which is estimated to have risen from full-glacial ELA of 2,830 m to altitudes of above 3,000 m. The icecaps on the plateau probably deglaciated by general downwasting rather than recession, for they had no high source areas towards which they could recede.

An estimate of the relative output from each source area is given in column H for full-glacial conditions and in column I for waning glacial conditions. These values are calculated by multiplying the area of each source by the relative thickness of accumulation under both full-glacial and waning glacial conditions. The point of the calculation is to be able to approximate the percentage contribution of each source to the total. For full-glacial conditions (column H), these percentages show that the Beartooth uplift provided about half and the icecaps on the plateau (II, III) provided about a third of the ice of the northern Yellowstone glacier. Under waning glacial conditions (column I) the Beartooth uplift provided about three-fourths of the ice, with this relative increase being largely due to more rapid recession of the icecaps on the plateau.

The "incursion factor" (column J) is the ratio of the percent values in column I to those in column H. Under waning glacial conditions, ice from the source with the higher value in column J will tend to move into the full-glacial domain of adjacent source areas that have lower values. Thus ice from the Gallatin Range and from the Beartooth uplift (factors 1.2 and 1.4, column J) will tend to move into the full-glacial domain of the plateau icecaps (factors of 0.5, column J), as indeed it did, most notably during the Deckard Flats readjustment. Ice from the Beartooth uplift will also tend to move into the full-glacial domain of ice in the upper Lamar drainage basin, as it did in late Pinedale time (pl. 3).

Figure 51 is a plot of the ice-volume changes of various source areas through time. The glacial cycle is considerably simplified and is based on relations during the Pinedale Glaciation. A sine-type curve is assumed for purposes of presentation. The two small perturbations shown near the end of the curve represent the Deckard Flats readjustment and the late Pinedale advance or standstill. Actual history probably involves more oscillations, and the buildup curve probably did not have the same slopes as the deglacial curve.

The vertical scale of figure 51 is the percent of the full-glacial volume of each icemass relative to *itself* only; the size of the different ice masses relative to one another is a different factor (table 8, columns H and I). Much of the information for the plot is based on field observation of the incursions of ice into the areas previously occupied by ice from another source.

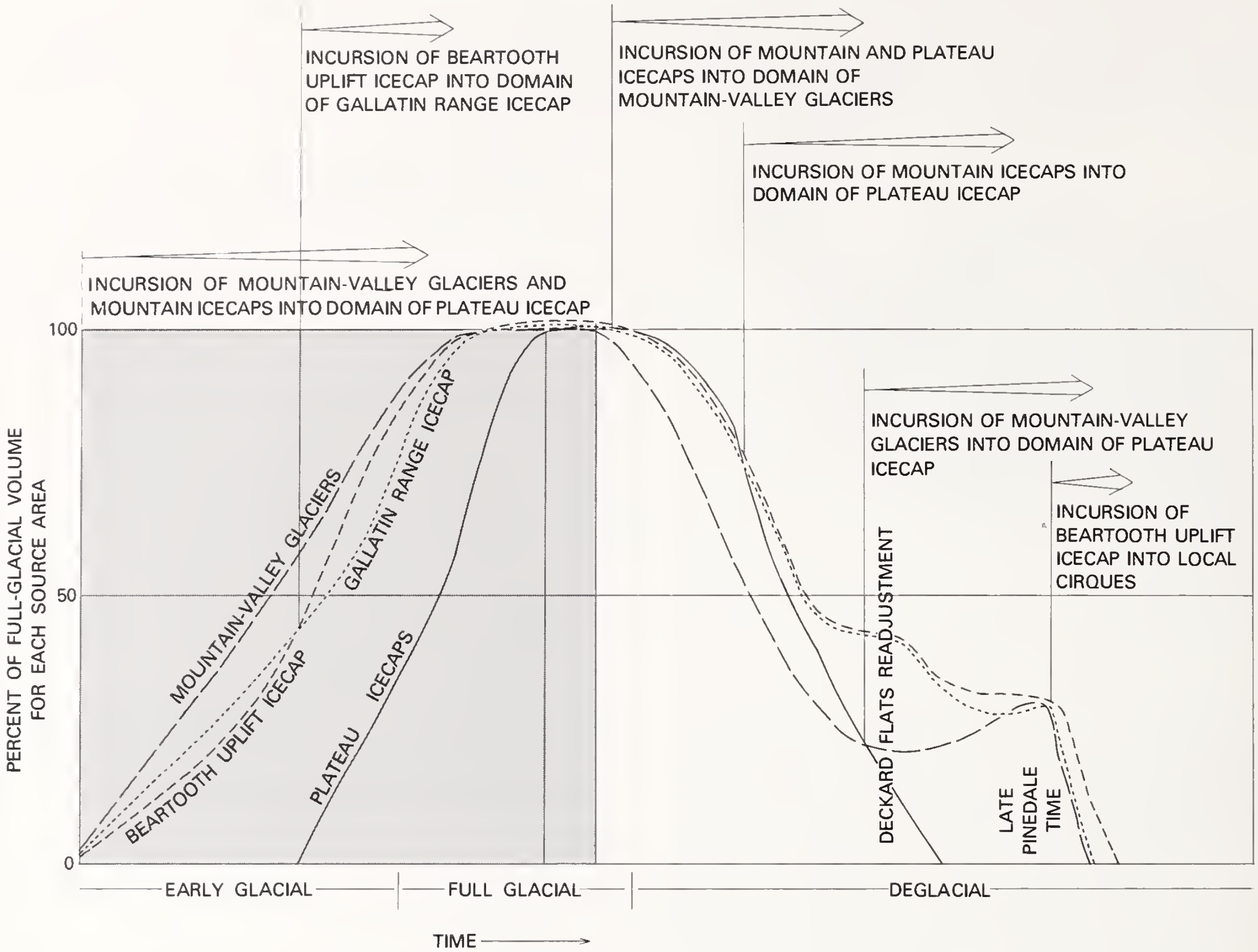


FIGURE 51.—Inferred phase relations between the several ice masses in the northern Yellowstone area. The vertical axis is for each ice mass relative to its full-glacial volume. The point where one curve crosses another is determined from field observations of incursions of ice from one source area into the full-glacial domain of another source area, as noted by the labeled arrows above the curves. The plateau icecap had a stronger "peaking" response than the other ice sources, as indicated by the narrower base of this curve. Left half of the diagram is shaded to emphasize paucity of field relations to control interrelations of curves there.

In summary one can draw the following conclusions from figure 51: (1) the plateau icecaps had stronger "peaking" responses than the other ice sources; (2) the larger mountain and plateau icecaps grew toward maximal values more slowly but remained longer near maximal ice volumes than the smaller mountain-valley glaciers; (3) the icecaps built on higher terrain, such as the Beartooth and Gallatin icecaps, both started earlier and remained longer than the icecaps built on lower terrain, such as the plateau icecaps; and (4) the relative change in volume of ice among the different source areas varied during both glacial buildup and recession; this lack of simultaneous phase relations indicates that terminal and recessional moraines deposited by glaciers from different kinds of source areas are likely not to be of exactly the same age.

REFERENCES CITED

Alden, W. C., 1928, Glaciation of Yellowstone National Park and its environs: Ranger Naturalists' Manual of Yellowstone National Park, mimeographed, p. 58–70. Available Yellowstone Naturalist's office.

______1932, Physiography and glacial geology of eastern Montana and adjacent areas: U.S. Geol. Survey Prof. Paper 174, 133 p.

______1953, Physiography and glacial geology of western Montana and adjacent areas: U.S. Geol. Survey Prof. Paper 231, 200 p.

Andrews, J. T., 1975, Glacial systems: North Scituate, Mass., Duxbury Press, 191 p.

Baker, R. G., 1970, Pollen sequence from late Quaternary sediments in Yellowstone Park: Science, v. 168, p. 1449–1450.

______1976, Late Quaternary vegetation history of the Yellowstone Lake Basin, Wyoming: U.S. Geol. Survey Prof. Paper 729–E, 48 p.

Baker, V. R., 1973, Paleohydrology and sedimentology of Lake Missoula flooding in eastern Washington: Geol. Soc. America Spec. Paper 144, 79 p.

Birkeland, P. W., Crandell, D. R., and Richmond, G. M., 1971, Status of correlation of Quaternary stratigraphic units in the western conterminous United States: Quaternary Research, v. 1, no. 2, p. 208-227.

Birkeland, P. W., and Sroba, R. R., 1974, The status of the concept of Quaternary soil-forming intervals in the western United States, *in* Mahaney, W. C., ed., Quaternary environments—Proceedings of a symposium: Toronto, Geographical Monographs, York University-Atkinson College, p. 241-276.

Blackwelder, Eliot, 1915, Post-Cretaceous history of the mountains in central western Wyoming: Jour. Geology, v. 23, no. 4, p. 97-117, 193-217, 307-340.

Boyd, F. R., 1961, Welded tuffs and flows in the rhyolite plateau of Yellowstone Park, Wyoming: Geol. Soc. America Bull., v. 72, no. 3, p. 387-426.

Bretz, J. H., Smith, H. T. U., and Neff, G. E., 1956, Channeled scabland of Washington—new data and interpretations: Geol. Soc. America Bull., v. 67, no. 8, p. 957-1049.

Bright, R. C., 1966, Pollen and seed stratigraphy of Swan Lake, southeastern Idaho: Tebiwa, v. 9, p. 1-47.

Broecker, W. S., Ewing, M., and Heezen, B. C., 1960, Evidence for an abrupt change in climate close to 11,000 years ago: Am. Jour. Sci., v. 258, p. 429-448.

Broecker, W. S., and van Donk, J., 1970, Insolation changes, ice volumes, and the O^{18} record in deep-sea cores; Rev. of Geophysics and Space Physics, v. 8, no. 1, p. 169-198.

Brown, C. W., 1961, Cenozoic stratigraphy and structural geology, northeast Yellowstone National Park, Wyoming and Montana: Geol. Soc. America Bull., v. 72, no. 8, p. 1173-1194.

Budd, W. F., 1975, A first simple model for periodically self-surging glaciers: Jour. Glaciology, v. 14, p. 3-21.

Budd, W. F., and Jenssen, D., 1975, Numerical modeling of glacier systems: Int. Assoc. Sci. Hyd., pub. 104, p. 257-291.

Charlesworth, J. K., 1957, The Quaternary Era: London, Edward Arnold Ltd., v. 1, 591 p.

Christiansen, R. L., and Blank, H. R., Jr., 1972, Volcanic stratigraphy of the Quaternary rhyolite plateau in Yellowstone National Park: U.S. Geol. Survey Prof. Paper 729-B, 18 p.

Clague, J. J., 1975, Glacier-flow patterns and the origin of late Wisconsinan till in the Southern Rocky Mountain Trench, British Columbia: Geol. Soc. America Bull., v. 86, p. 721-731.

Clark, M. M., 1967, Pleistocene glaciation of the drainage of the West Walker River, Sierra Nevada, California: Stanford, Calif., Stanford Univ., Ph. D. thesis, 130 p.

Clayton, Lee, 1967, Stagnant-glacier features of the Missouri Coteau in North Dakota, *in* Glacial geology of the Missouri Coteau and adjacent areas: North Dakota Geol. Survey Misc. Ser. 30, p. 25-46.

Colman, S. M., 1977, The development of weathering rinds on basalts and andesites and their use as a Quaternary dating method, western United States: University of Colorado Ph.D. dissertation, 235 p.

Crandell, D. R., 1965, The glacial history of western Washington and Oregon, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, N. J., Princeton Univ. Press, p. 341-353.

Crittenden, M. D., Jr., 1963, New data on the isostatic deformation of Lake Bonneville: U.S. Geol. Survey Prof. Paper 454-E, 31 p.

Dirks, R. A., 1974, Climatological studies of Yellowstone and Grand Teton National Parks: Laramie, Wyo., Univ. Wyoming, Dept. Atmospheric Science, 37 p.

Dreimanis, A., 1953, Studies of friction cracks along shores of Cirrus Lake and Kasakokwog Lake, Ontario. Am. Jour. Sci., v. 251, p. 769-783.

Dreimanis, A., and Karrow, P. F., 1972, Glacial history in the Great Lakes-St. Lawrence region, the classification of the Wisconsin(an) Stage, and its correlatives, *in* Quaternary geology: Internat. Geol. Cong., 24th, Montreal, 1972, sec. 12, p. 5-15.

Emiliani, Cesare, 1966, Paleotemperature analysis of Caribbean cores P6304-8 and P6304-9 and a generalized temperature curve for the past 425,000 years: Jour. Geology, v. 74, no. 2, p. 109-126.

______1972, Quaternary paleotemperatures and the duration of the high-temperature intervals: Science, v. 178, p. 398-401.

Farnes, P. E., and Shafer, B. A., 1975, Summary of snow survey measurements for Montana, 1922-1974: Bozeman, Mont., U.S. Soil Conservation Service, 220 p.

Flint, R. F., 1955, Pleistocene geology of eastern South Dakota: U.S. Geol. Survey Prof. Paper 262, 173 p.

______1971, Glacial and Quaternary geology: New York, John Wiley and Sons, 892 p.

Foose, R. M., Wise, D. U., and Garbarini, G. S., 1961, Structural geology of the Beartooth Mountains, Montana and Wyoming: Geol. Soc. America Bull., v. 72, p. 1143-1172.

Fournier, R. O., White, D. E., and Truesdell, A. H., 1970, Discharge of thermal water and heat from Upper, Midway, and Lower Geyser Basins, Yellowstone Park [abs.]: American Quaternary Assoc., First Meeting, p. 47.

Fraser, G. D., Waldrop, H. A., and Hyden, H. J., 1969, Geology of the Gardiner area, Park County, Montana: U.S. Geol. Survey Bull. 1277, 118 p.

Friedman, Irving, 1968, Hydration rind dates rhyolite flows: Science, v. 159, p. 878-879.

Friedman, Irving, and Long, William, 1976: Hydration rate of obsidian: Science, v. 191, p. 347-352.

Friedman, Irving, Pierce, K. L., Obradovich, J. D., and Long, W. D., 1973, Obsidian hydration dates glacial loading?: Science, v. 180, p. 733-734.

Friedman, Irving, and Smith, R. L., 1960, A new dating method using obsidian— Part I, the development of the method: Am. Antiquity, v. 25, p. 476-522.

Fryxell, Roald, 1965, Mazama and Glacier Peak volcanic ash layers: Relative ages: Science, v. 147, p. 1288-1290.

Fryxell, Roald, Richmond, G. M., and Malde, H. E., 1965, Pullman to Boise, *in* Guidebook for Field Conference E, northern and middle Rocky Mountains, Internat. Assoc. Quaternary Research, 7th Congr.: Lincoln, Nebraska, Acad. Sci., p. 90-91.

Gadd, N. R., 1971, Pleistocene geology of the central St. Lawrence lowland: Canada Geol. Survey Mem. 359, 153 p.

Geikie, Archibald, 1881, The geysirs of the Yellowstone: MacMillians Magazine, October, 1881, p. 1-16.

Gilbert, G. K., 1906, Crescentic gouges on glaciated surfaces: Geol. Soc. America Bull., v. 17, p. 303-314.

Glen, J. W., 1955, The creep of polycrystalline ice: Roy. Soc. London Proc. ser. A, v. 228, p. 519-538.

Good, J. M., 1965, Prehistoric landslide in Yankee Jim Canyon, Park County, Montana [abs.]: Geol. Soc. America Spec. Paper 82, p. 327-328.

______1964, Relict Wisconsin ice in Yellowstone National Park, Wyoming: Contributions to Geology, Univ. of Wyoming, v. 3, p. 92-93.

Gravenor, C. P., 1955, The origin and significance of prairie mounds: Am. Jour. Sci., v. 253, p. 475-481.

Hague, Arnold, 1904, Atlas to accompany Monograph 32 on the geology of the Yellowstone National Park: U.S. Geol. Survey Mon. 32, 37 sheets.

______1912, Geological history of the Yellowstone National Park: U.S. Dept. Interior, 24 p.

Hague, Arnold, Iddings, J. P., Weed, W. H., Walcott, C. D., Girty, G. H., Stanton, T. W., and Knowlton, F. H., 1899, Descriptive geology, petrography, and paleontology, Pt. 2 *of* Geology of Yellowstone National Park: U.S. Geol. Survey Mon. 32, 893 p.

Hall, W. B., 1960, Multiple glaciation in the Madison and Gallatin Ranges, southwestern Montana, *in* West Yellowstone–Earthquake area, Billings Geol. Soc. Guidebook 11th Ann. Field Conf., 1960: p. 191–206.

______1961, Geology of the upper part of the Gallatin valley of southwestern Montana: University of Wyoming Ph. D. thesis.

Hamilton, W. B., 1960, Late Cenozoic tectonics and volcanism of the Yellowstone Region, Wyoming, Montana, and Idaho, *in* West Yellowstone–Earthquake area, Billings Geol. Soc. Guidebook 11th Ann. Field Conf., 1960: p. 92–105.

______1964, Volcanic rocks of the West Yellowstone and Madison Junction quadrangles, Montana, Wyoming, and Idaho: U.S. Geol. Survey Prof. Paper 435–S, p. 209–221.

Harris, S. E., Jr., 1943, Friction cracks and the direction of glacial movement: Jour. Geology, v. 51, p. 244–258.

Heusser, C. J., 1965, A Pleistocene phytogeographical sketch of the Pacific Northwest and Alaska, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, N. J., Princeton Univ. Press, p. 469–483.

Hodge, S. M., 1972, The movement and basal sliding of the Nisqually Glacier, Mount Rainier: Scientific Report no. 13, Department of Atmospheric Sciences, Univ. of Washington, Seattle

Holmes, W. H., 1881, Glacial phenomena in the Yellowstone Park: American Naturalist, v. 15, p. 203–208.

______1883, Report on the geology of the Yellowstone National Park, *in* Hayden, F. V., A report of progress of the exploration in Wyoming and Idaho for the year 1878: U.S. Geol. and Geog. Survey of the Terr., 12th Ann. Rept., pt. 2, p. 1–87.

Horberg, Leland, 1940, Geomorphic problems and glacial geology of the Yellowstone Valley, Park County, Montana: Jour. Geology, v. 48, no. 3, p. 275–303.

Howard, A. D., 1937, History of the Grand Canyon of the Yellowstone: Geol. Soc. America Spec. Paper 6, 159 p.

Landsberg, Helmut, 1958, Physical climatology: Dubois, Pa., Gray Printing Co., 446 p.

Leaf, C. F., 1975, Watershed management in the Rocky Mountain subalpine zone— the status of our knowledge: U.S. Forest Service Research Paper RM–137, 31 p.

Love, J. D., 1961, Reconnaissance study of Quaternary faults in and south of Yellowstone National Park, Wyoming: Geol. Soc. America Bull., v. 72, no. 12, p. 1749–1764.

Love, J. D., and Reed, J. C. Jr., 1971, Creation of the Teton landscape: Grand Teton Natural History Association, Moose, Wyoming, 120 p.

Machette, M. N., Birkeland, P. W., Markos, G., and Guccione, M. J., 1976a, Soil development in Quaternary deposits in the Golden-Boulder portion of the Colorado Piedmont, *in* Epis, R. C., and Weimer, R. J., eds., Studies in Colorado field geology: Prof. Contr. Colorado School Mines, no. 8, p. 339–357.

Machette, M. N., Birkeland, P. W., Burke, R. M., Guccione, M. J., Kihl, Rolf, and Markos, Gergely, 1976b, Field descriptions and laboratory data for a Quaternary soil sequence in the Golden-Boulder portion of the Colorado Piedmont: U.S. Geol. Survey Open-File Rept. 76–804, 21 p.

Malde, H. E., 1968, The catastrophic late Pleistocene Bonneville flood in the Snake River Plain, Idaho: U.S. Geol. Survey Prof. Paper 596, 52 p.

Mathews, W. H., 1967, Profiles of late Pleistocene glaciers in New Zealand: New Zealand Jour. Geol. and Geophys., v. 10, p. 146–163.

______1974, Surface profiles of the Laurentide ice sheet in its marginal areas: Jour. Glaciology, v. 13, p. 37–43.

McKee, E. D., Crosby, E. J., and Berryhill, H. L., Jr., 1967, Flood deposits, Bijou Creek, Colorado, June, 1965: Jour. Sed. Pet., v. 37, p. 829–851.

Meier, M. F., 1960, Mode of flow of the Saskatchewan Glacier, Alberta, Canada: U.S. Geol. Survey Prof. Paper 351, 70 p.

______1968, Calculations of slip of Nisqually Glacier on its bed—No simple relation of sliding velocity to shear stress: Int. Assoc. Sci. Hyd. pub. 79, p. 49–57.

Meier, M. F., Kamb, W. B., Allen, C. R., and Sharp, R. P., 1974, Flow of Blue Glacier, Olympic Mountains, Washington, U.S.A.: Jour. Glaciology, v. 13, p. 187-212.

Meier, M. F., and Post, Austin, 1962, Recent variations in mass net budgets of glaciers in western North America: Int. Assoc. Sci. Hyd. pub. 58, p. 63–77.

Meier, M. F., Tangborn, W. V., Mayo, L. R., and Post, Austin, 1971, Combined ice and water balances of Gulkana and Wolverine Glaciers, Alaska, and South Cascade Glacier, Washington, 1965 and 1966 hydrologic years: U.S. Geol. Survey Prof. Paper 715–A, 23 p.

Mercanton, P. L., 1916, Vermessungen am Rhonegletscher, Neue Denkschrifter der Schweiz: Naturforschenden Gesellschaft, v. 52, 190 p.

Miner, N. A., 1937, Pleistocene glaciation of the Gardiner, Mammoth Hot Springs, and Lava Creek regions, Yellowstone National Park, Wyoming: Iowa City, Iowa, Iowa St. Univ., Ph. D. thesis.

Montagne, John, 1968, Geomorphology and Cenozoic history of the Yellowstone valley south of Livingston, Montana: Geol. Soc. America, Rocky Mountain Sec. Guidebook, Field Trip no. 5, 9 p.

______1970, Highlights of the Quaternary geomorphic history of Yellowstone valley between Gardiner and Livingston, Montana [abs.]: American Quaternary Assoc., Abstract supplement, First Meeting, p. 14–15.

Morrison, R. B., 1965, Quaternary geology of the Great Basin, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, N.J., Princeton Univ. Press, p. 265–285.

Mudge, M. R., 1967, Surficial geologic map of the Sawtooth Ridge quadrangle, Teton and Lewis and Clark Counties, Montana: U.S. Geol. Survey Geol. Quad. Map GQ–610.

Mullineaux, D. R., Hyde, J. H., and Rubin, Meyer, 1972, Preliminary assessment of upper Pleistocene and Holocene pumiceous tephra from Mount St. Helens, southern Washington: Geol. Soc. America Abs. with Programs, v. 4, no. 3, p. 204–205.

Naeser, C. W., Izett, G. A., and Wilcox, R. E., 1973, Zircon fission-track ages of the Pearlette family ash beds in Meade County, Kansas: Geology, v. 1, p. 187–189.

Nye, J. F., 1952, The mechanics of glacier flow: Jour. Glaciology, v. 2, p. 82–93.

______1953, The flow law of ice from measurements in glacier tunnels, laboratory experiments and the Jungfraufirm borehole experiment: Roy. Soc. London Proc. ser. A, v. 219, p. 477–489.

______1959, The motion of ice sheets and glaciers: Jour. Glaciology, v. 3, p. 493–507.

______1965, The flow of a glacier in a channel of rectangular, elliptic, or parabolic cross section: Jour. Glaciology, v. 5, p. 661–690.

Okko, Veikko, 1950, Friction cracks in Finland: Soc. Geol. de Finlande, Comptes Rendus, no. 23, p. 45–51.

Østrem, Gunnar, 1966a, The height of the glaciation limit in southern British Columbia and Alberta: Geografiska Annaler, v. 48A, p. 126–138.

______1966b, Mass balance studies on glaciers in western Canada, 1965: Ottawa, Ontario, Canada Dept. Mines and Tech. Surveys, Geog. Br. Geographical Bull., v. 8, p. 81–107.

Østrem, Gunnar, and Liestol, Olav, 1964, Glasiologiske undersøkelser i Norge 1963: Norsk Polarinstitutt Meddelelser nr. 90, p. 281–340.

Paterson, W. S. B., 1969, The physics of glaciers: New York, Pergamon Press, 250 p.

______1970, The sliding velocity of Athabasca Glacier, Canada: Jour. Glaciology, v. 9, p. 55–63.

Peak, G. W., and Clagett, G. P., 1973, Summary of snow survey measurements, Wyoming, and pertinent measurements in Colorado, Montana, and South Dakota, 1919–1972: Casper, Wyo., U.S. Soil Conservation Service, 206 p.

Pierce, K. L., 1968, Evidence of north-flowing, pre-Pinedale glaciation, crest of the Washburn Range, Yellowstone National Park [abs.]: Geol. Soc. America Spec. Paper no. 121, p. 621.

______1973a, Surficial geologic map of the Mount Holmes quadrangle and parts of the Tepee Creek, Crown Butte, and Miner quadrangles, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Misc. Geol. Inv. Map I–640.

______1973b, Surficial geologic map of the Mammoth quadrangle and part of the Gardiner quadrangle, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Misc. Geol. Inv. Map I–641.

______1974a, Surficial geologic map of the Tower Junction quadrangle and part of the Mount Wallace quadrangle, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Misc. Geol. Inv. Map I–647.

______1974b, Surficial geologic map of the Abiathar Peak quadrangle, and parts of the Cutoff Mountain, Pilot Peak, and Sunlight Peak quadrangles, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Misc. Geol. Inv. Map I–646.

Pierce, K. L., Obradovich, J. D., and Friedman, Irving, 1976, Obsidian hydration dating and correlation of Bull Lake and Pinedale Glaciations near West Yellowstone, Montana: Geol. Soc. America Bull., v. 87, p. 703–710.

Pierce, W. G., 1965a, Geologic map of the Clark quadrangle, Park County, Wyoming: U.S. Geol. Survey Geol. Quad. Map GQ–477.

______1965b, Geologic map of the Deep Lake quadrangle: U.S. Geol. Survey Geol. Quad Map GQ–478.

Pierce, W. G., and Nelson, W. H., 1969, Geologic map of the Wapiti quadrangle, Park County, Wyoming: U.S. Geol. Survey Geol. Quad. Map GQ–778.

Porter, S. C., 1975, Equilibrium-line altitudes of late Quaternary glaciers in the southern Alps, New Zealand: Quaternary Research, v. 5, p. 27–47.

Prostka, H. J., Ruppel, E. T., and Christiansen, R. L., 1975, Geologic map of the Abiathar Peak quadrangle, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Geol. Quad. Map GQ–1244.

Prostka, H. J., Blank, H. R., Jr., Christiansen, R. L., and Ruppel, E. T., 1975, Geologic map of the Tower Junction quadrangle, Yellowstone National Park, Wyoming and Montana: U.S. Geol. Survey Geol. Quad. Map GQ–1247.

Pytte, Randi, and Østrem, Gunnar, 1965, Glasio-hydrologiske undersøkelser i. Norge 1964: Norges Vassdrags- og Elektrisitetsvesen, meddelelse fra Hydrologisk Avdeling. nr. 14, 92 p.

Richmond, G. M., 1962a, Quaternary stratigraphy of the La Sal Mountains, Utah: U.S. Geol. Survey Prof. Paper 324, 135 p.

______1962b, Three pre-Bull Lake tills in the Wind River Mountains, Wyoming, *in* Geological Survey research 1962: U.S. Geol. Survey Prof. Paper 450–D, p. D132–D136.

______1964a, Glacial geology of the West Yellowstone Basin and adjacent parts of Yellowstone National Park: U.S. Geol. Survey Prof. Paper 435–T, p. 223–236.

______1964b, Three pre-Bull Lake tills in the Wind River Mountains, Wyoming—a reinterpretation, *in* Geological Survey research 1964: U.S. Geol. Survey Prof. Paper 501–D, p. D104–D109.

______1965a, Glaciation of the Rocky Mountains, *in* Wright, H. E., Jr. and Frey, D. G., eds., The Quaternary of the United States: Princeton Univ. Press, p. 217–230.

______1965b, Relation of Madison Plateau flow to Bull Lake moraine, *in* Guidebook for Field Conference E, northern and middle Rocky Mountains, Internat. Assoc. Quat. Res., 7th Cong.: Lincoln, Nebr., Nebraska Acad. Sci., p. 48–49.

______1968, Late Pinedale and Neoglacial glaciation in the Absaroka Range, east of Yellowstone Lake, Wyoming [abs.]: Geol. Soc. America Spec. Paper 121, p. 627–628.

______1969, Development and stagnation of the last Pleistocene icecap in the Yellowstone Lake Basin, Yellowstone National Park, U.S.A.: Eiszeitalter u. Gegenwart, Band. 20, p. 196–203.

______1970a, Comparison of the Quaternary stratigraphy of the Alps and Rocky Mountains: Quaternary Research, v. 1, no. 1, p. 3–28.

______1970b, Glacial history of the Yellowstone Lake Basin [abs]: American Quaternary Assoc., First Meeting, p. 112–113.

______1972, Appraisal of the future climate of the Holocene in the Rocky Mountains: Quaternary Research, v. 2, p. 315–322.

______1973a, Surficial geologic map of the Warm River Butte quadrangle, Yellowstone National Park and adjoining areas, Idaho and Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–645.

______1973b, Surficial geologic map of the Grassy Lake Reservoir quadrangle, Yellowstone National Park and adjoining area, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–644.

______1973c, Surficial geologic map of the West Thumb quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–643.

______1973d, Surficial geologic map of the Huckleberry Mountain quadrangle: U.S. Geol. Survey Misc. Geol. Inv. Map I–639.

______1974, Surficial geologic map of the Frank Island quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–642.

______1976, Pleistocene stratigraphy and chronology in the mountains of western Wyoming, *in* Mahaney, W. C., ed., Quaternary stratigraphy of North America: Stroudsburg, Pa., Dowden, Hutchinson and Ross, Inc., p. 353–379.

______1977a, Surficial geologic map of the Canyon Village quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I–652.

______1977b, Extensive glaciers in Yellowstone National Park ~114,000 and ~88,000 years ago, Abstracts, X INQUA Congress, Birmingham, p. 382.

Richmond, G. M., and Hamilton, W. B. 1960, The late Quaternary age of obsidian-rhyolite flows in the western part of Yellowstone National Park, Wyoming, *in* Geological Survey research 1960: U.S. Geol. Survey Prof. Paper 400–B, p. B224–B225.

Richmond, G. M., and Murphy, J. F., 1965, Geologic map of the Bull Lake East quadrangle, Fremont County, Wyoming: U.S. Geol. Survey Geol. Quad. Map GQ–431.

Richmond, G. M., and Pierce, K. L., 1968, Late Pleistocene center of ice-accumulation over Yellowstone Lake area [abs.]: Geol. Soc. America Spec. Paper 115, p. 441.

______1971a, Surficial geologic map of the Mount Hancock quad-

rangle, Yellowstone National Park and adjoining area, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-636.

______1971b, Surficial geologic map of the Two Ocean Pass quadrangle, Yellowstone National Park and adjoining area, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-635.

______1972, Surficial geologic map of the Eagle Peak quadrangle, Yellowstone National Park and adjoining area, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-637.

Richmond, G. M., and Waldrop, H. A., 1972, Surficial geologic map of the Pelican Cone quadrangle, Yellowstone National Park and adjoining area, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-638.

______1975, Surficial geologic map of the Norris Junction quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-650.

Ritter, D. F., 1967, Terrace development along the front of the Beartooth Mountains, southern Montana: Geol. Soc. America Bull., v. 78, p. 467-484.

Ross, C. P., Andrews, D. A., and Witkind, I. J., 1955, Geologic map of Montana: U.S. Geol. Survey.

Ruppel, E. T., 1972, Geology of pre-Tertiary rocks in the northern part of Yellowstone National Park: U.S. Geol. Survey Prof. Paper 729-A, 66 p.

Schmidt, D. L., and Mackin, J. H., 1970, Quaternary geology of Long and Bear Valleys, west-central Idaho: U.S. Geol. Survey Bull. 1311-A, 22 p.

Schul'tz, V. L., ed., 1962, Lednik Fedchenko: Academiia nauk Uzbekskeĭ SSR, Tashkent, v. 2, 198 p.

Schytt, Valter, 1967 A study of "ablation gradient": Geografiska Annaler, v. 49A, p. 327-332.

Scott, G. R., 1965, Nonglacial Quaternary geology of the southern and middle Rocky Mountains, *in* Wright, H. E., Jr. and Frey, D. G., eds., The Quaternary of the United States: Princeton, N.J., Princeton Univ. Press, p. 243-254.

Seager, G. F. 1944, Gold, arsenic, and tungsten deposits of the Jardine-Crevasse Mountain district, Park County, Montana: Montana Bureau of Mines and Geology Mem. no. 23, 111 p.

Shackleton, N. J., and Opdyke, N. D., 1973, Oxygen isotope and palaeomagnetic stratigraphy of Equatorial Pacific core V28-238—Oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale: Quaternary Research, v. 3, no. 1, p. 39-55.

Shreve, R. L., and Sharp, R. P., 1970, Internal deformation and thermal anomalies in the lower Blue Glacier, Mount Olympus, Washington, USA: Journal of Glaciology, v. 9, p. 65-86.

Shroba, R. R., 1976, Soil development on Pinedale, Bull Lake, and pre-Bull Lake tills, Sawatch Range and Front Range, Colorado, Sangre de Cristo Range, New Mexico, and Wind River Range, Wyoming: Geol. Soc. America Abs. with Programs, v. 8, no. 5, p. 629.

Smedes, H. W., and Prostka, H. J., 1972, Stratigraphic framework of the Absaroka Volcanic Supergroup in the Yellowstone National Park region: U.S. Geol. Survey Prof. Paper 729-C, 33 p.

Steiger, R. H., and Jager, E., 1977, Convention on the use of decay constants in geo- and cosmochronology: Earth and Planetary Science Letters, v. 35, p. 359-362.

Tangborn, W. V., Mayo, L. R., Scully, D. R., and Krimmel, R. M., 1977, Combined ice and water balances of Maclure Glacier, California, South Cascade Glacier, Washington, and Wolverine and Gulkana Glaciers, Alaska, 1967 Hydrologic Year: U.S. Geol. Survey Prof. Paper 715-B, 20 p.

Trimble, D. E., 1963, Geology of Portland, Oregon, and adjacent areas: U.S. Geol. Survey Bull. 1119, 119 p.

Trimble, D. E., and Carr, W. J., 1961, The Michaud delta and the Bonneville River near Pocatello, Idaho, *in* Short papers in the geologic and hydrologic sciences: U.S. Geol. Survey Prof. Paper 424-B, p. B164-B166.

U.S. Geological Survey, 1972a, Surficial geologic map of Yellowstone National Park: U.S. Geol. Survey Misc. Geol. Inv. Map I-710.

______1972b, Geologic map of Yellowstone National Park: U.S. Geol. Survey Misc. Geol. Inv. Map I-711.

______1974, Surface water supply of the United States, 1966-70, Missouri River Basin above Williston, North Dakota: U.S. Geol. Survey Water Supply Paper 2116, Part 6, v. 1, 835 p.

U.S. Weather Bureau, 1934-7, Climatic summary of the United States from the establishment of the stations to 1930, inclusive: sections 6, 9, 11, 12, and 13.

______1960, Climates of the states: nos. 60-10, 60-24, and 48.

Virkkala, Kalevi, 1960, On the striations and glacier movements in the Tempere region, southern Finland: Soc. Geol. de Finlande, Comptes Rendus, no. 32, p. 159-177.

Waddington, J. C. B., and Wright, H. E., Jr., 1974, Late Quaternary vegetational changes on the east side of Yellowstone Park, Wyoming: Quaternary Research, v. 4, p. 175-184.

Wahrhaftig, Clyde, and Cox, Allan, 1959, Rock glaciers in the Alaska Range: Geol. Soc. America Bull., v. 70, p. 383-436.

Waldrop, H. A., 1970, Glaciation of west-central Yellowstone National Park [abs.]: American Quaternary Assoc. First Meeting, p. 142-143.

______1975a, Surficial geologic map of the West Yellowstone quadrangle, Yellowstone National Park and adjoining area, Wyoming, Idaho, and Montana: U.S. Geol. Survey Misc. Geol. Inv. Map I-648.

______1975b, Surficial geologic map of the Old Faithful Quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-649.

Waldrop, H. A., and Hyden, H. J., 1963, Landslides near Gardiner, Montana: U.S. Geol. Survey Prof. Paper 450-E, p. E11-E14.

Waldrop, H. A., and Pierce, K. L., 1975, Surficial geologic map of the Madison Junction quadrangle, Yellowstone National Park, Wyoming: U.S. Geol. Survey Misc. Geol. Inv. Map I-651.

Wedow, Helmuth, Jr., Gaskill, D. L., Banister, D. P., and Pattee, E. C., 1975, Mineral resources of the Absaroka primitive area and vicinity, Park and Sweet Grass Counties, Montana: U.S. Geol. Survey Bull. 1391-B, 115 p.

Weed, W. H., 1893, The glaciation of the Yellowstone Valley north of the Park: U.S. Geol. Survey Bull. 104, 41 p.

Witkind, I. J., 1969, Geology of the Tepee Creek quadrangle, Montana-Wyoming: U.S. Geol. Survey Prof. Paper 609, 101 p.

INDEX

[Page numbers of major references are in italic]

UNITED STATES DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

EXPLANATION

Area covered by ice during the Deckard Flats readjustment—Stippling indicates areas of end moraines, thick till, and ice-contact deposits of Deckard Flats age. Areas covered by ice outside of Yellowstone National Park are not shown on this plate. Outline dashed where approximately located, queried where uncertain

Glacial striation formed by recessional Pinedale ice

Flow direction indicated by glacial scour, streaming, or molding feature of recessional Pinedale ice

Approximate line of transport of glacial erratics

Inferred thickness, in feet, and flowdirection of Deckard Flats ice

Crests of Chico moraines and moraines backfilling Tom Miner Basin

Ice dam and lake sediments in tributary valley

Southern limit of abundant erratics of Precambrian rocks from the Beartooth uplift. Emplaced during Pinedale recession

Isolated erratics of Precambrian rocks from the Beartooth uplift—Probably carried southward prior to Pinedale full-glacial time, may have been later reworked (See fig 14)

Conversion of feet to meters
1 ft = 0.3048 m

feet	meters	feet	meters
100	30	5,000	1,524
200	61	6,000	1,829
300	91	7,000	2,134
400	122	8,000	2,438
500	152	9,000	2,743
600	183	10,000	3,048
700	213	11,000	3,353
800	244	12,000	3,658
900	274		

This map shows the area occupied by Deckard Flats ice in the northern part of Yellowstone National Park. The type area is east of Gardiner, Montana. A younger Deckard Flats phase is represented by moraines southeast of Mammoth Hot Springs—the Gardiner River Bridge moraines. Direction of movement of Deckard Flats ice was locally 60°–130° different from that under full-glacial conditions (pl. 1*A*). Differences were greatest where glaciers from icecaps on the Gallatin Range and Beartooth uplift moved into areas occupied under full-glacial conditions by the icecaps on the Yellowstone plateau

North of the park several recessional features older than Deckard Flats indicate that the northern Yellowstone outlet glacier and not glaciers from local tributary valleys provided most of the ice to the outlet glacier. In addition to the glacial flow features alined parallel to the Yellowstone valley (pl 1*A*), the evidence showing that the outlet glacier was dominant over local tributary glaciers is: (1) morainal backfills, such as in the Tom Miner Basin area, (2) ice-dammed lake sediments in the tributary valleys, as along Sixmile and Emigrant Creeks, and (3) flow lines defined by transport of erratics at high levels across the mouths of tributary valleys, such as above Mol Heron and Bear Creeks.

Another recessional feature older than the Deckard Flats readjustment is shown on the northwest flank of the Washburn Range. Just below an altitude of 2,700 m (8,800 ft), striations and erratics of Precambrian rocks from the Beartooth uplift indicate that southwestward flow occurred after ice levels had diminished so that northwest flow across the range crest was cut off.

Base from AMS, Ashton, Idaho, Mont. Wyo., 1955, Billings, Mont. Wyo., 1954 Bozeman, Mont. Wyo. 1958 Cody, Wyo. 1955

SCALE 1:250 000

5 0 5 10 15 MILES

5 0 5 10 15 KILOMETERS

CONTOUR INTERVAL 200 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929
1979 MAGNETIC DECLINATION AT SOUTH EDGE VARIES FROM 16°15' TO 15°30' EAST

Based on field work, 1965–1974, mostly shown on Pierce (1973a, b, and 1974a, b). Mapping of the Chico moraine modified from Horberg (1940, fig 3) and Montagne (1970). Distribution of ice-dammed lake sediments along Emigrant and Sixmile Creeks modified from Montagne (1970). The line of transport of erratics across the valley of Mol Heron Creek modified from Weed (1893, pl 1)

DECKARD FLATS READJUSTMENT AND OTHER PINEDALE RECESSIONAL FEATURES, NORTHERN YELLOWSTONE NATIONAL PARK AREA, MONTANA AND WYOMING

U.S. GOVERNMENT PRINTING OFFICE: 1979-677-129/36

Base from AMS, Ashton, Idaho, Mont., Wyo., 1955, Billings, Mont., Wyo., 1954 Bozeman, Mont., Wyo., 1958 Cody, Wyo., 1955

SCALE 1:250 000

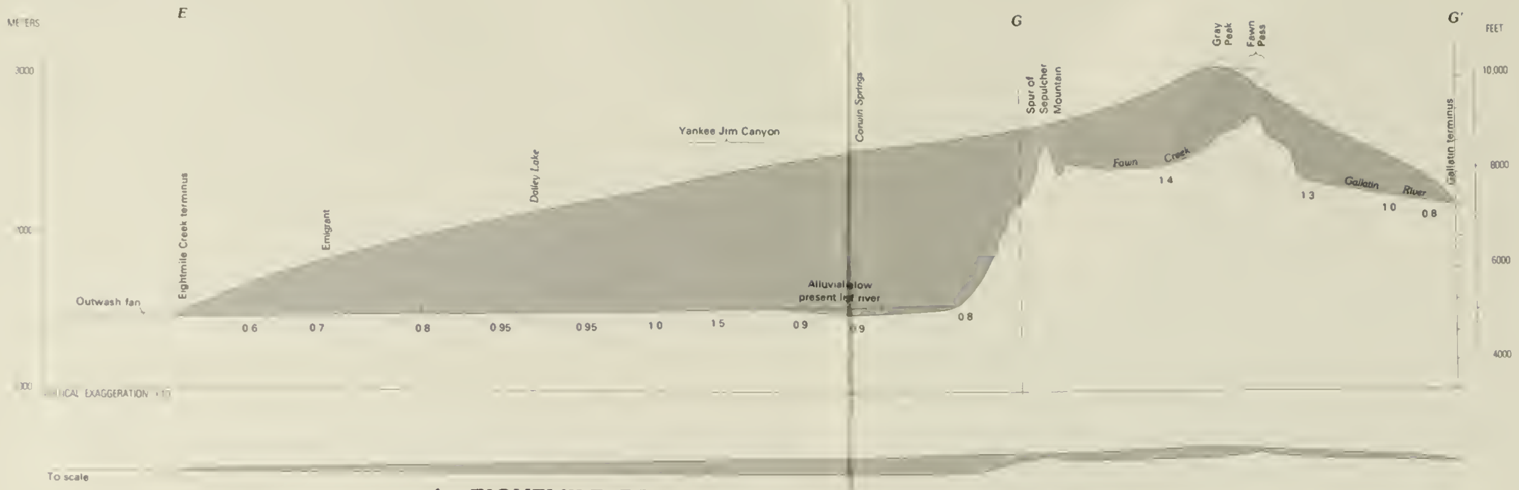
5 0 5 10 15 MILES

5 0 5 10 15 KILOMETERS

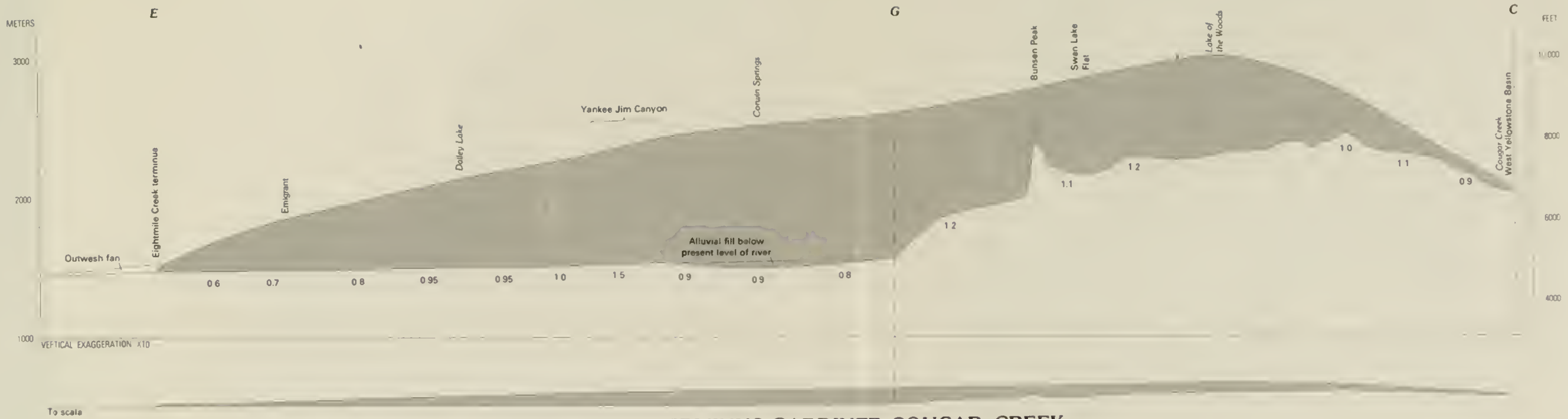
CONTOUR INTERVAL 200 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929
1979 MAGNETIC DECLINATION AT SOUTH EDGE VARIES FROM 16°15' TO 15°30' EAST

EXTENT OF LATE PINEDALE GLACIERS AND LATE PINEDALE AND YOUNGER FLOODS, NORTHERN YELLOWSTONE NATIONAL PARK AREA, MONTANA AND WYOMING

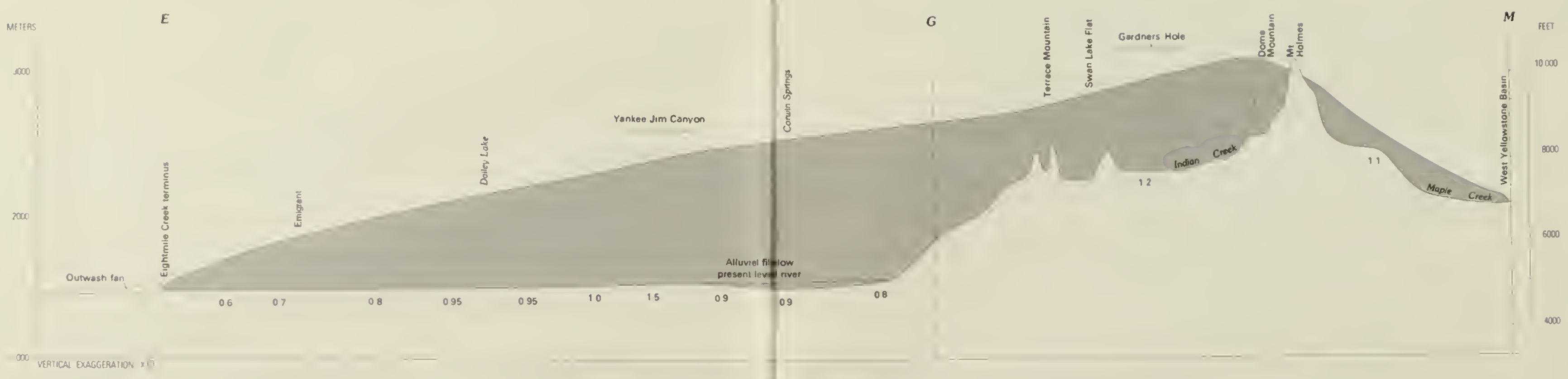
☆U.S. GOVERNMENT PRINTING OFFICE: 1979–677-129/26



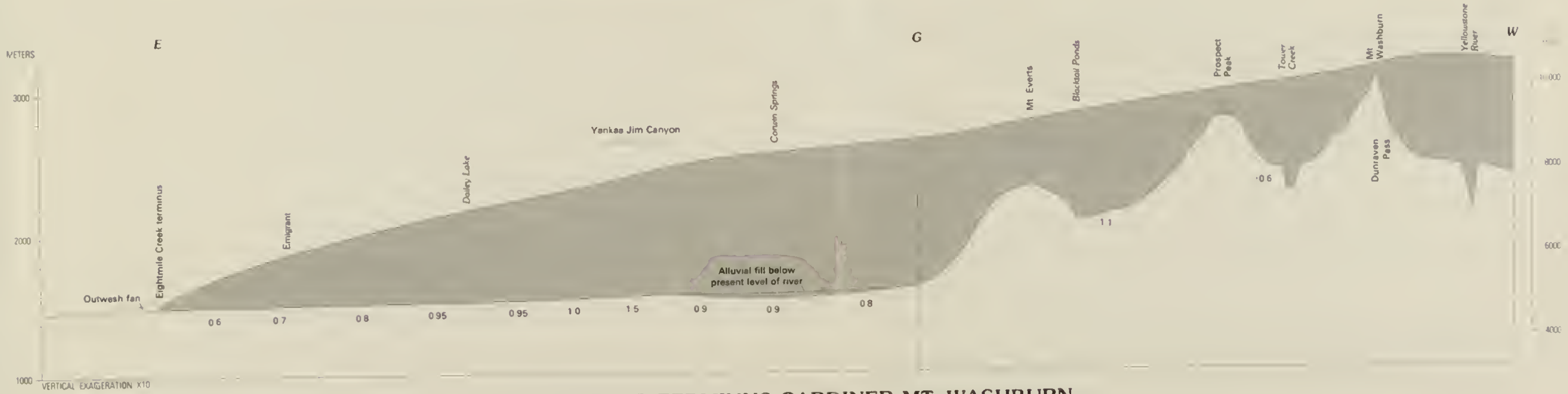
A—EIGHTMILE TERMINUS-GARDINER-GALLATIN RIVER



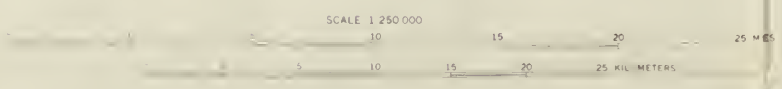
B—EIGHTMILE TERMINUS-GARDINER-COUGAR CREEK



C—EIGHTMILE TERMINUS-GARDINER-MAPLE CREEK

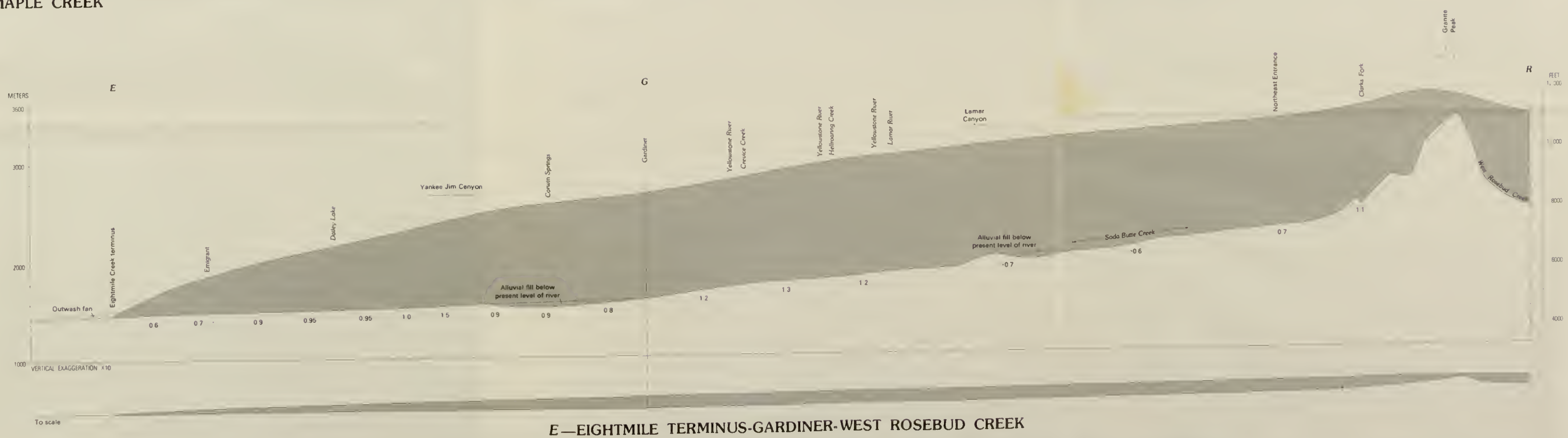


D—EIGHTMILE TERMINUS-GARDINER-MT. WASHBURN



The locations of the sections are shown on plate 1A. Patterned area indicates configuration of full-glacial icemass. The sections extend from the Pinedale terminus in the Yellowstone valley, up to Gardiner, from which sections are drawn towards the various icecap sources and then down to termini in (1) the Gallatin River valley, (2) the Maple Creek valley, and (3) the Cougar Creek valley.

The profile sections show that the thicker the ice, the more gentle the slope of the ice surface. The cosecant (1/sin) of the surface slope tends to be proportional to the thickness, as indicated by the equation for primary basal shear stress in text of report. Beneath the profile sections are shown the values of basal shear stress as calculated from the ice slope, thickness, and valley shape. (See fig 48, and table 3, this report.) In some places the inferred values of ice slope and thickness indicate values of basal shear stress that seem too low, such as the Washburn Range (sec. EGW) and the Lamar valley (sec. EGR). There the altitude shown for the ice surface is the minimum altitude provided by the highest peaks; the actual altitude of the ice surface was probably 100 m higher. Short-dashed lines, projection into plane of sections of nearby geographic features.



E—EIGHTMILE TERMINUS-GARDINER-WEST ROSEBUD CREEK

PROFILES AND SECTIONS OF FULL-GLACIAL PINEDALE ICEMASS, NORTHERN YELLOWSTONE NATIONAL PARK AREA, MONTANA AND WYOMING

EXPLANATION

Glacier surface above and below inferred equilibrium-line altitude (ELA)

Contours on surface of ice, in thousands of feet—Solid where surface of ice is defined on nearby nunataks or valley walls long dashed where approximately located or where altitude of contour is only several hundred feet above freshly glaciated uplands short dashed where deduced from flow pattern and where altitude of contour is well above the high glaciated uplands Contours are perpendicular to full-glacial flow features See chart for conversion to meters

Direction of flow and glacier thickness in feet based on ice surface contours—See chart for conversion to meters

Direction of glacier flow—Solid arrow, observed in field and on aerial photographs broken arrow observed only on aerial photographs Indicated by glacial streaming, molding, or scouring features Where full-glacial and deglacial flow directions differed, only full-glacial flow features are shown

Glacial striation—Arrows indicate direction of ice movement

Glacial outwash channels

Ice divide—Arrows indicate flow away from divide

Approximate line of transport of glacial erratics

Subdivisions of the northern Yellowstone glacier based on icecap source areas—I, Gallatin Range icecap, II, plateau icecap between the Washburn and Gallatin Ranges, III, plateau icecap flowing across Washburn Range Specimen Ridge area, IV, Beartooth uplift icecap, V, upper Lamar icecap

Present main drainage divides

Lines of profile sections (pl 4)

Most of the glacial ice in northern Yellowstone was part of the northern Yellowstone glacier, defined as the outlet glacier in the Yellowstone valley and the ice that fed it The source area of the northern Yellowstone glacier was from four interconnected icecaps ice from these sources is delimited by roman numerals I to V This ice converged towards Gardiner, Montana, where it formed the 55 km long outlet glacier The northern Yellowstone glacier headed along an ice-divide whose shape was rectangular in plan view excepting a missing section occupied by the outlet glacier at the position of the northwest corner of this rectangle This ice divide did not correspond precisely with the present drainage divides, but was offset by as much as several kilometers Glacial scour features along the present drainage divides indicate that sizable glaciers flowed across these divides from the Yellowstone River drainage For example, ice flowed west across the drainage divide of the southern Gallatin Range, where it joined glaciers from local cirques and flowed to terminal moraines in the valleys of the Gallatin River and Grayling Creek

Conversion of feet to meters
1 ft = 0 3048 m

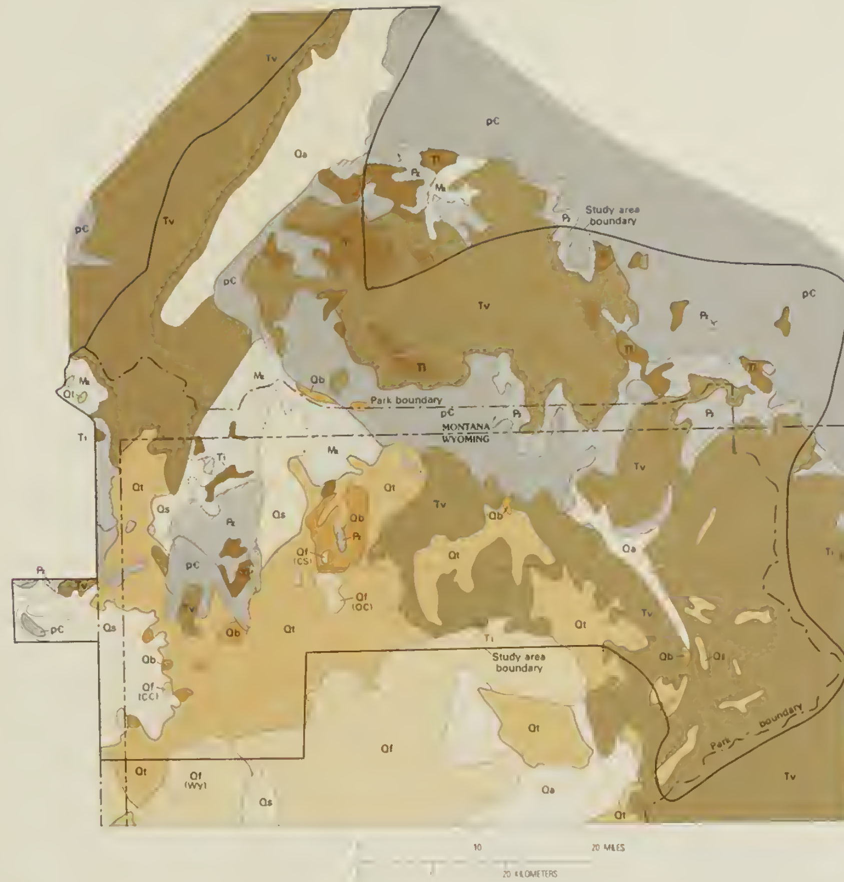
feet	meters	feet	meters
100	30	5,000	1,524
200	61	6,000	1,829
300	91	7,000	2,134
400	122	8,000	2,438
500	152	9,000	2,743
600	183	10,000	3,048
700	213	11,000	3,353
800	244	12,000	3,658
900	274		

Base from AMS Ashton Idaho Mont Wyo 1955 Billings Mont Wyo 1954 Bozeman Mont Wyo 1958 Cody Wyo 1956

SCALE 1 250 000
5 0 5 10 15 MILES
5 0 5 10 15 KILOMETERS
CONTOUR INTERVAL 200 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929
1979 MAGNETIC DECLINATION AT SOUTH EDGE VARIES FROM 16°15′ TO 15°30′ EAST

Glacial geology based on fieldwork between 1965–1974 mostly shown on Pierce (1973a b and 1974a b) Madison Canyon area based on Waldrop and Pierce (1975), and glacial geology along east park boundary south of Frost Pass modified from Richmond and Waldrop (1972) Limit of Pinedale glaciers from Corwin Springs north along the Yellowstone valley modified from Weed (1893) Horberg (1940) Montagne (1968 1970 written commun 1972) and my own field observations Limit of Pinedale glaciers and direction of scour features near drainage divide of Beartooth uplift based on reconnaissance field studies and on more than 450 areas of glacial scour noted on aerial photographs

A—PINEDALE FULL-GLACIAL ICEMASS



EXPLANATION

Qa	Thick Quaternary surficial deposits
Qb	Quaternary basalt
Qf	Quaternary rhyolite flows—CC, Cougar Creek flow, CS, Crystal Springs flow, OC Obsidian Cliff flow, WY, West Yellowstone flow
Qt	Quaternary rhyolite tuff
Tv	Tertiary Absaroka Volcanic Supergroup (andesitic)
Ti	Tertiary intrusive rocks
Ms	Mesozoic shale and sandstone
Pz	Paleozoic limestone, shale, and sandstone
pC	Precambrian granitic rocks and schist

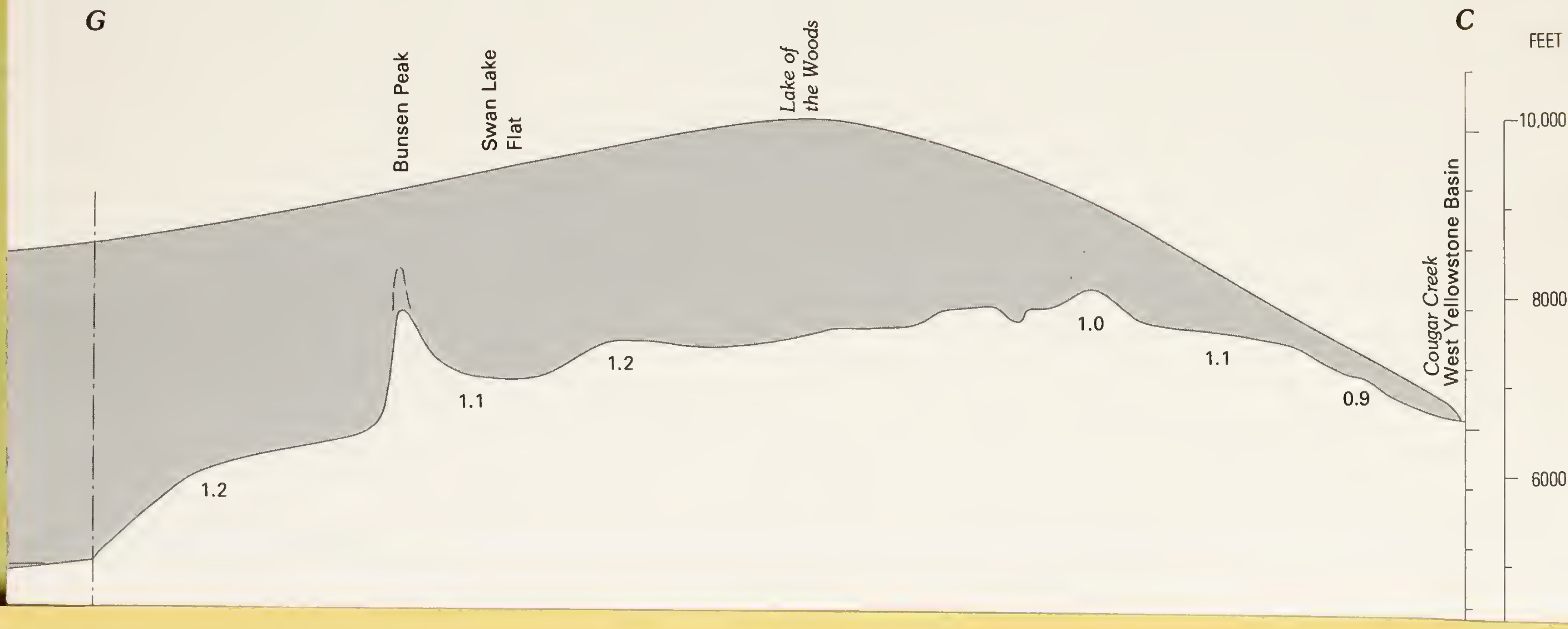
The Quaternary tuff and Tertiary volcanic rocks that underlie most of the area do not record or preserve evidence of glaciation very well This accounts for the lack of recognition of widespread glaciation in previous studies Sketch map generalized after U S Geol Survey (1972b) and a compilation by Ross, Andrews, and Witkind (1955)

B—MAJOR BEDROCK TYPES

INDEX MAP SHOWING LOCATION OF STUDY AREA

PINEDALE FULL-GLACIAL ICEMASS AND MAJOR BEDROCK TYPES IN THE NORTHERN YELLOWSTONE NATIONAL PARK AREA, WYOMING AND MONTANA

☆ U.S. GOVERNMENT PRINTING OFFICE: 1980–677-129/26


G
C
FEET
Bunsen Peak
Swan Lake
Flat
Lake of
the Woods
Cougar Creek
West Yellowstone Basin
1.2
1.1
1.2
1.0
1.1
0.9
10,000
8000
6000

1879 · CENTENNIAL · 1979
U.S. GEOLOGICAL SURVEY